Reconnaissance Stratigraphy of the Prichard Formation (Middle Proterozoic) and the Early Development of the Belt Basin, Washington, Idaho, and Montana

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[Plates are in pocket]

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[For readers who wish to convert measurements from the metric system of units to U.S. customary units, the conversion factors are listed below]

Metric unit	Multiply by	To obtain U.S. customary unit
micrometer (μm)	3.937×10^{-5}	inch
millimeter (mm)	3.937×10^{-2}	inch
centimeter (cm)	3.937×10^{-1}	inch
meter (m)	3.281	foot
kilometer (km)	6.214×10^{-1}	mile
kilometer^2 (km^2)	3.861×10^{-1}	mile^2
kilometer^3 (km^3)	2.399×10^{-1}	mile^3
kilogram (kg)	2.205	pound
gram/centimeter^3 (g/cm^3)	6.243×10	pound/foot^3

RECONNAISSANCE STRATIGRAPHY OF THE PRICHARD FORMATION (MIDDLE PROTEROZOIC) AND THE EARLY DEVELOPMENT OF THE BELT BASIN, WASHINGTON, IDAHO, AND MONTANA

By EARLE R. CRESSMAN

ABSTRACT

The Prichard Formation crops out in the area between Glacier National Park on the east and Spokane, Washington, on the west and from the United States–Canada border on the north nearly to the Idaho batholith on the south. The formation, commonly more than 6 kilometers thick, comprises the basal one-fourth to one-third of the Belt Supergroup, and the base is everywhere concealed.

The Prichard differs from the rest of the Belt Supergroup in containing a great thickness of turbidite, abundant and widespread diorite sills, laminated carbon-rich silts, and abundant iron sulfide. The rocks, which have been regionally metamorphosed to the greenschist facies, are within the biotite to garnet zone everywhere except in Glacier National Park. Original metamorphism was static and resulted from load, but Mesozoic compression and intrusion superposed dynamic metamorphism on the rocks in the western and southwestern parts of the outcrop area.

Rocks of the Prichard Formation, which are classified as quartzite, siltite, and argillite, consist mostly of quartz, plagioclase (mostly albite), sericite (2M illite), biotite, and chlorite. The quartzites are graywackes. The sills are mostly gabbro to quartz diorite, and their composition is characteristic of continental tholeiite.

The Prichard Formation consists of an argillaceous facies, present in the St. Joe, Coeur d'Alene, and Salish Mountains in the southern and eastern parts of the area, and a quartzite facies present in the Purcell, northern Cabinet, and Selkirk Mountains. Each facies is divided into informal members for purposes of mapping, and although the nomenclature is clumsy and not entirely in accord with the North American Stratigraphic Code, it is used in this report.

The argillaceous facies includes, from base to top, members A through H and the transition member. Members A and B, totaling 1,600 meters thick, consist of graded siltite-argillite couplets less than one to several centimeters thick that probably were deposited from low-density turbidity currents. Slump folds, common in the uppermost 1,000 meters, indicate an east-facing slope. Member C, about 100 meters thick, is mostly argillic quartzites that were deposited as turbidites. Sole marks and ripple marks indicate a southerly sediment transport direction, and the sediment source probably lay east of the depositional basin. Member D is mostly olive-gray platy-weathering argillite that becomes more silty upsection. It is 290 meters thick in the eastern Coeur d'Alene Mountains but less than 100 meters thick in the western Coeur d'Alenes. Member E consists of nearly 1,000 meters of interlaminated siltite and argillite interbedded with some well-sorted crossbedded quartzite. The unit contains shallow-water to intertidal features that range from hummocky cross-stratification in the lower and middle parts to desiccation cracks near the top.

That part of the argillite facies above member E is more argillic and less silty than the underlying members. Member F is mostly planar interlaminated and interbedded argillite and silty argillite that contain abundant iron sulfide laminae. These beds probably represent both fine-grained turbidites and hemipelagic deposits. Debris-flow deposits are present locally near the base. The contact with member E below is sharp at most localities. In the northeastern part of the report area the interval that elsewhere is occupied by argillites of member F consists of dolomitic siltite, probably derived from a carbonate shelf on the east side of the depositional basin and transported by turbidity currents. Member G consists of interbedded quartzite turbidite, siltite turbidite, and interlaminated siltite and argillite. The unit is a tongue of the quartzite member of the quartzite facies and ranges in thickness from a feather edge to more than 1,000 meters. Member H, like member F, consists mostly of planar-laminated and interbedded silty argillite and argillite. Near Plains, Montana, member H is about 1,500 meters thick, but elsewhere it ranges from about 600 to 750 meters thick. The transition member rests in sharp contact on member H and is about 800 meters thick throughout most of the area. The transition member, which is a shelf deposit, consists of a basal massive silty argillite, a middle unit of irregularly laminated siltite and argillite that makes up most of the member, and an upper unit of interbedded quartzite and irregularly interbedded siltite and argillite. The contact with the overlying Burke Formation, placed at the top of the uppermost unit of interlaminated light-gray and dark-gray argillite and siltite, is difficult to locate consistently.

The quartzite facies is made up mostly of the quartzite member, which consists of intervals 2–45 meters thick of predominant quartzite that alternate with intervals 2–20 meters thick of argillite and siltite. Quartzite makes up two-thirds of the member. The quartzite beds, mostly 0.2–0.4 meter thick, are turbidites; flute, groove, and ridge-and-furrow casts are present though not abundant. The siltite and most of the argillite were deposited from low-density turbidity currents, though some of the argillite beds are hemipelagic deposits of wide extent.

The massive member of the quartzite facies, as much as 250 meters thick, is present only in the western Purcell Mountains where it is intercalated in the lower part of the quartzite member. The massive member, which displays no bedding, probably resulted from fluidization of a sand-silt-clay section as a result of sill intrusion and the consequent conversion of pore water to steam.

Both the argillite member and the upper member of the quartzite facies are probably tongues of member H of the argillaceous facies. The argillite member is about 240 meters thick and occurs intercalated in the upper part of the quartzite member about 1,600 meters below the top of the Prichard Formation. The upper member, present throughout the area of the quartzite facies, is about 400–600 meters thick and is overlain by the transition member, which is common to both facies.

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The maximum age of the Prichard Formation is 1,600 million years, which is the minimum age of the pre-Belt basement, and the minimum age is 1,330±45 million years, which is the time of formation of metamorphic biotite near the top of the Prichard. A sill within the formation yielded a uranium-lead zircon age of 1,433±30 million years. This date probably applies to about the middle of the quartzite member. Assuming the subsidence rate to have been a direct function of the square root of time and calibrating the subsidence curve of the supergroup by two published ages, one of which is speculative, it is calculated that the quartzite facies may have been deposited through a time span of 36 million years, which ended 1,410 million years ago.

The Prichard Formation represents two progradational sequences; the first consists of members A through E of the argillaceous facies, and the second consists of the quartzite facies and member F through the transition member of the argillaceous facies. The sediments of both sequences were contributed by large rivers. The source of the lower sequence was somewhere west of the outcrop areas; sediments of the upper sequence were contributed by a major river that entered the area from the south or southwest. The sediments of the upper sequence formed a submarine fan comparable in size to that of the Pliocene-Pleistocene fan off the mouth of the Mississippi River. Thicknesses and transport directions indicate that the apex of the fan was in the southwest part of the report area. Turbidity currents that transported the sediments were deflected to the left by the Coriolis force, confirming paleomagnetic data that the area was in the southern hemisphere. The drainage area that supplied the vast amount of terrigenous material to the Prichard Formation probably included not only the southern part of Middle Proterozoic North America but also parts of the Siberian and other shields that were assembled into a supercontinent.

The Belt basin seems to have formed by divergence between the adjacent North American and Siberian shields that stretched the intervening continental crust. Stretching was probably accompanied by the intrusion of basic magma and was followed by subsidence. The subsidence formed a gulf that opened northward to the world ocean. The two progradational sequences of the Prichard Formation resulted from two episodes of stretching, basin formation, and basin fill. At the end of Belt time the two shields converged. The development of the Belt basin followed a pattern common in Proterozoic time.

INTRODUCTION

PURPOSE AND SCOPE

The Belt Supergroup and its Canadian equivalent, the Purcell Supergroup, crop out throughout an area of about 130,000 km^2 in the northwestern United States and western Canada (fig. 1). The Belt is locally as much as 16 km thick (Harrison and others, 1986), the base is not exposed, and the top is eroded. It is thus the thickest and one of the most extensive accumulations of Proterozoic rocks of sedimentary origin.

The Belt Supergroup consists of four major units. These are, from the base up, the informally designated lower Belt, the Ravalli Group, the middle Belt carbonate (Helena and Wallace Formations), and the Missoula Group. In the area of this report (fig. 1) the lower Belt comprises a single formation, the Prichard, which constitutes the basal one-fourth to one-third of the entire Belt Supergroup. The Prichard Formation is the subject of this report.

The Prichard Formation differs from the rest of the Belt Supergroup in several respects. The Prichard contains laminated carbon-rich argillites and thick marine turbidite sequences, whereas most of the rest of the Belt contains abundant evidence of deposition in shallow-marine and subaerial environments. The Prichard was intruded by abundant and widespread basic sills, whereas such sills, though present locally, are uncommon in the rest of the supergroup. Iron in the Prichard is mostly in sulfides, but in the rest of the Belt (with some important exceptions) the iron is mostly in oxides. Furthermore, the Aldridge Formation, British Columbia's equivalent of the Prichard Formation, contains the Sullivan lead-zinc-silver deposit that is stratiform and syngenetic.

In this report I present stratigraphic sections of the Prichard Formation throughout its area of occurrence, describe its lithologic character, infer depositional environments and paleogeography, and speculate on the early history of the Belt basin. The area considered is shaded on figure 1.

This study is reconnaissance in nature. Because of the paucity of previous studies, the scarcity of detailed geologic maps, and the great thickness and extent of the Prichard Formation, it could hardly be otherwise.

PREVIOUS STUDIES

The Prichard Formation was first described in the Coeur d'Alene mining district by Ransome (1905, p. 279–281) and was named the "Prichard slate", though the credit for describing and naming the formation apparently belongs to Frank C. Calkins (Ransome, 1905, p. 274, 275). The name was derived from Prichard Creek, a tributary of the Coeur d'Alene River in the northern part of the Coeur d'Alene mining district (fig. 2). Ransome (1905, p. 28) described the Prichard as follows:

> The Prichard slate is the thickest and the most homogeneous of the formations in the Coeur d'Alene district and occupies the greatest area. It is also one of the most distinctive, the regularly banded bluish-gray slates being readily recognized. It can be distinguished from certain somewhat similar beds in the Wallace Formation by its noncalcareous character and by the fact that it weathers in reddish-brown tints, while the weathered exposures of the Wallace formation are yellowish gray.

The formation was described as follows in a tabular section also on page 28:

> Mostly blue-black, blue-gray to light gray slates, generally distinctly banded. Considerable interbedded gray sandstone. Upper portion characterized by rapid alternations of argillaceous and arenaceous layers, and by shallow-water features. Base not exposed.

The thickness was given as about 8,000 ft (2,500 m). An expanded description was given by Ransome and

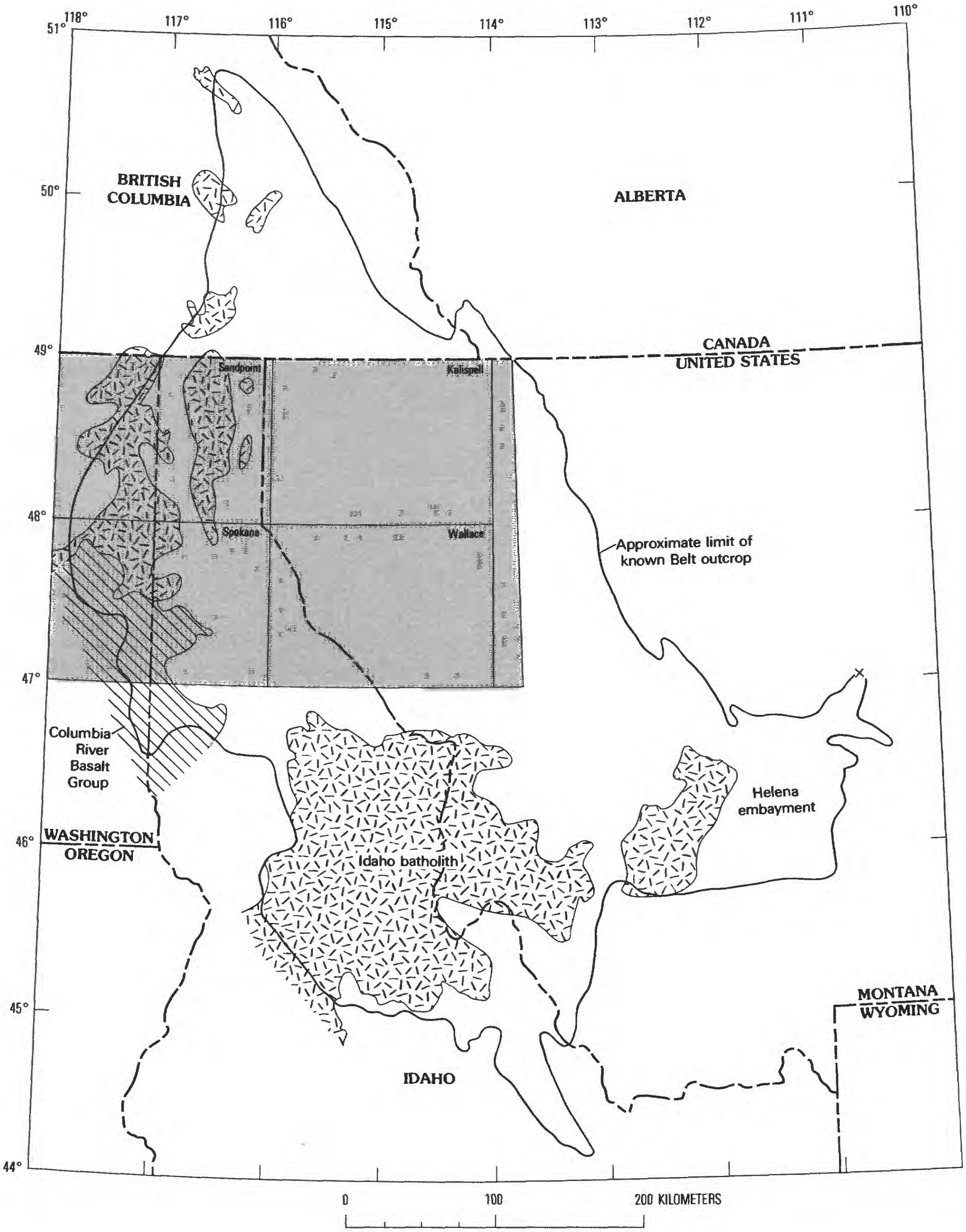


FIGURE 1.—Extent of Belt Supergroup (United States) and Purcell Supergroup (Canada). Shows area considered in this report (shaded), batholithic terrane (patterned), Columbia River Basalt Group (lined), and names of 1°×2° quadrangles. Exposures in Little Belt Mountains of Montana where Belt Supergroup rests in sedimentary contact on Archean rocks marked by x.

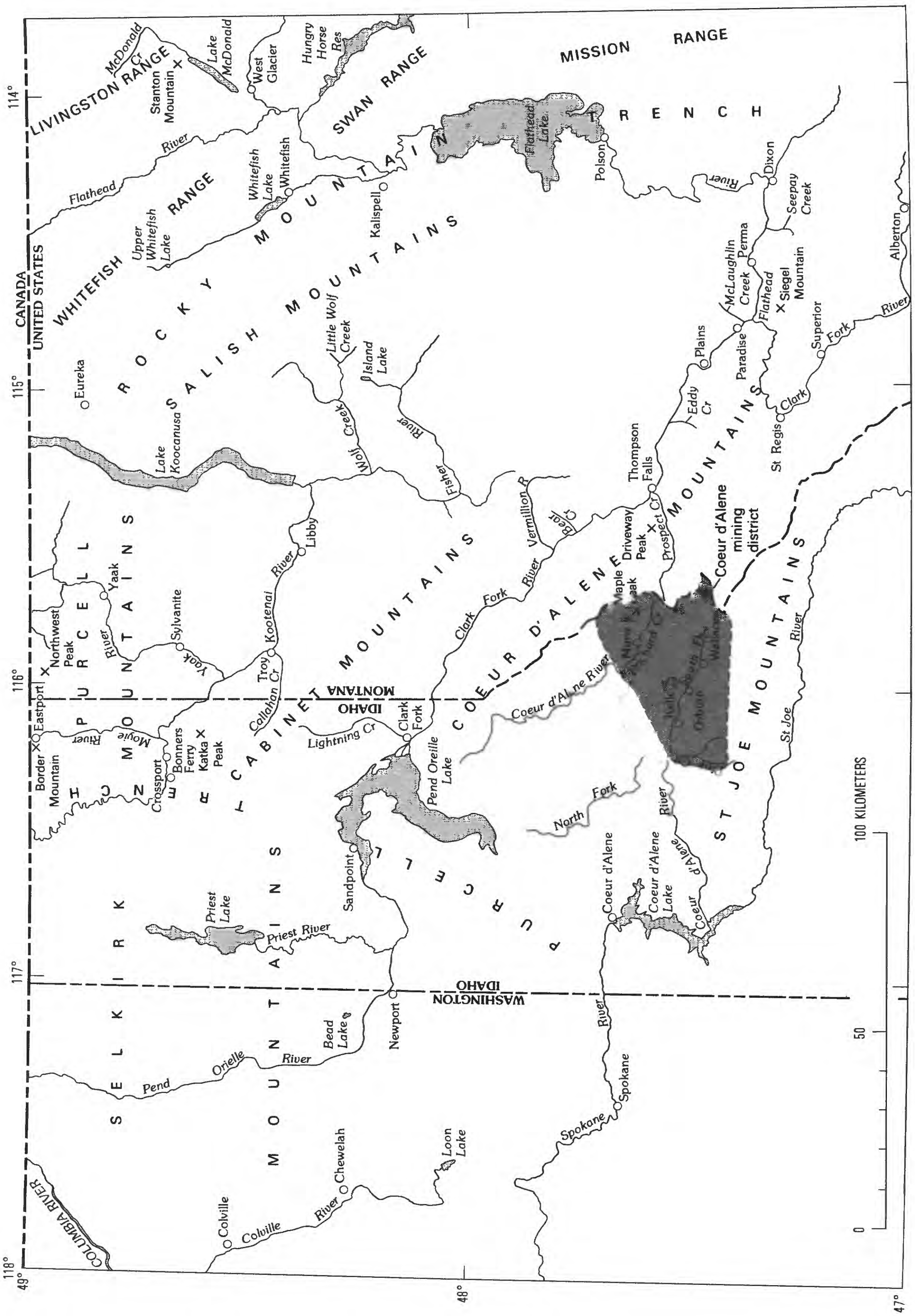


FIGURE 2.—Map showing principal geographic features of area of report.

Calkins (1908, p. 23, 24, 29–32) in their final report on the Coeur d'Alene district.

Calkins (1909) conducted a remarkable reconnaissance of northern Idaho and northwestern Montana in 1905 in order to assess the extent of the formations in the Belt Supergroup that had been identified in the Coeur d'Alene district. Calkins identified the Prichard Formation as far east as the Flathead River near Perma, Mont., as far west as Coeur d'Alene Lake, and as far north as the United States–Canada border. He noted (Calkins, 1909, p. 36–37) that in the Coeur d'Alene mining district, near Vermillion Creek, Mont., and near Plains, Mont., the Prichard Formation is mostly banded argillite, whereas in the vicinity of Pend Oreille Lake, along the Kootenai and Yaak Rivers east of Bonners Ferry, Idaho, and in the vicinity of the 49th parallel the formation consists mostly of gray quartzite.

In 1912, Schofield (1914, p. 22), working in the Purcell Mountains of southeastern British Columbia, applied the name Aldridge Formation to the basal unit of the Purcell Supergroup, the Canadian equivalent of the Belt Supergroup. By comparing the Canadian section with that of the Coeur d'Alene district and from Calkins' mapping along the international border, Schofield correlated his Aldridge Formation with the Prichard Formation of northwestern Montana and northern Idaho.

In the following years the extent of the Prichard Formation was determined in more detail, mostly by reconnaissance and small-scale mapping. Some examples are mapping north of Pend Oreille Lake by Anderson (1930), mapping of the Libby 30-minute quadrangle in 1929–1934 by Gibson (1948), and the mapping in reconnaissance of Lincoln and Flathead Counties, Mont., by Johns (1970). Maps at a scale of 1:24,000 have been published of the Coeur d'Alene mining district (Hobbs and others, 1965) and at a scale of 1:62,500 of the Alberton 15-minute quadrangle, Mont. (Wells, 1974), the Clark Fork, Packsaddle Mountain, and Elmira 15-minute quadrangles north of Pend Oreille Lake (Harrison and Jobin, 1963, 1965; Harrison and Schmidt, 1971, respectively), the Newport 30-minute quadrangle, Wash. (Miller, 1974a, b, c, d), and the Chewelah–Loon Lake area east of Chewelah, Wash. (Miller and Clark, 1975). In the last two decades the earlier work was supplemented by new mapping to produce 1:250,000-scale maps that now cover most of the Belt basin (Griggs, 1973; Mudge and others, 1982; Harrison and others, 1983, 1986). As a result, the distribution of the Prichard Formation is, for the most part, well delineated.

Little progress was made in determining the internal stratigraphy of the Prichard Formation until after World War II when Hobbs and others (1965) restudied the Coeur d'Alene mining district. They described the Prichard as being a monotonous succession of quartzose argillite and argillite that contains several lighter colored quartzitic zones (Hobbs and others, 1965, p. 30–34). They described the thickness and distribution of these zones in considerable detail. Their upper quartzite zone is the transition member of the Prichard of this report, their middle quartzite zone is member G of this report, and their lower quartzite zone coincides approximately with member E.

In the more quartzitic section north of Clark Fork, Idaho, Harrison and Jobin (1963, p. 8–10, pls. 1, 2) divided the Prichard Formation into a basal member more than 15,000 ft (4,600 m) thick and dominated by quartzite, a laminated argillite member 2,000–2,200 ft (610–670 m) thick, and an upper argillite and siltite member 800–1,000 ft (240–300 m) thick. These correspond to the quartzite member, the upper member, and the transition member, respectively, of this report. Harrison and Campbell (1963, p. 1417) presented average modal mineralogic composition determined by X-ray diffraction of quartzite, siltite, and argillite of the Prichard near Clark Fork, Harrison and Grimes (1970) gave semiquantitative spectrographic analyses from the same area, and Maxwell and Hower (1967) studied the mineralogy of illite in samples from the Clark Fork area. In eastern Washington, Miller (1974a, b, c, d) and Miller and Clark (1975) described several units within the Prichard Formation but did not map them. Webster (1981) measured partial sections near Paradise, Mont., described and measured slump folds, and documented the presence of turbidites. Finally, Finch and Baldwin (1984) proposed a basin-wide chronocorrelation of the Prichard and Aldridge Formations.

In British Columbia, the Aldridge Formation has generally been divided into three parts. The lower Aldridge consists of thinly to irregularly bedded rusty-weathering argillite, the middle Aldridge consists of gray quartzite and thinly bedded siltite and argillite about 3,000 m thick, and the upper Aldridge consists of 300–400 m of thinly bedded rusty-weathering argillite and siltite (Høy, 1982, p. 137). The middle Aldridge Formation, which contains the world-class Sullivan lead-zinc-silver deposits in its lower part, has been the subject of much study, mostly by geologists of Cominco Ltd. which operates the Sullivan mine. Little of this work has been published, but Bishop and others (1970) reported that the gray quartzites of the middle Aldridge are turbidites and that the transport directions were from south to north. Geologists of Cominco discovered within the middle Aldridge 14 beds of finely laminated argillite that are markers of regional extent (Thompson and Panteleyev, 1976). These marker beds are correlated on the basis of the pattern of banding that results from differing amounts of carbonaceous matter. Huebschman (1973) traced one of the marker beds about

160 km from Clark Fork, Idaho, northward to near Cranbrook, British Columbia. Edmunds (1977a) presented abundant mineralogic and chemical data on 100 samples of the Aldridge. Høy (1978, 1979) mapped several units within the Aldridge Formation in the Hughes Range east of the Rocky Mountain trench in southeastern British Columbia, measured current directions, made environmental interpretations, and presented evidence of penecontemporaneous faulting. McMechan (1979, 1981) also presented evidence of penecontemporaneous faulting in southeastern British Columbia along deeply rooted block faults that developed during rifting.

PRESENT INVESTIGATION

The principal area considered in this report, shaded in figure 1, consists of the Sandpoint, Kalispell, Spokane, and Wallace 1°×2° quadrangles and a narrow selvage to the east. This area includes all of the identified exposures of the Prichard Formation except for some intensively metamorphosed rocks in thrust sheets immediately north of the Idaho batholith.[1]

This investigation was conducted largely as part of the U.S. Geological Survey's 1°×2° (1:250,000-scale) mapping program. In the summers of 1979 and 1980 exposures of the Prichard Formation between Plains and Perma, Mont., in the Wallace 1°×2° quadrangle were mapped at a scale of 1:62,500 (Cressman, 1981, 1982, 1985; Harrison and others, 1986). The Prichard of that area consists mostly of argillite and is similar to the formation in its type area in the Coeur d'Alene mining district. In the summers of 1981 and 1982 the thick exposures of the Prichard Formation on both sides of the Yaak River in the northwest corner of the Kalispell 1°×2° quadrangle were mapped, and the results were published at a scale of 1:48,000 (Cressman and Harrison, 1986) and 1:250,000 (Harrison and others, 1983. The Prichard there consists largely of gray quartzite turbidites interbedded with argillite and siltite. Gently dipping exposures of the upper part of the Prichard Formation east of Lake Koocanusa and the much thicker and structurally more complex exposures between Troy, Mont., and the 117th meridian were mapped in 1983 and 1984, and the 1983 results were published at 1:250,000 as part of the Kalispell 1°×2° geologic map (Harrison and others, 1983). The summer of 1985 and two weeks in 1986 were spent studying other major outcrop areas and in making more detailed observations in areas previously mapped.

Plate 1 shows the distribution of exposures of the Prichard Formation in the report area and the locations of the stratigraphic sections that are the basis of this report. The sections are mostly more than 6 km thick, the northern part of the area is largely covered by glacial drift, most of the area is heavily forested, and the structure commonly is complex. It is therefore impossible to measure lengthy, continuous stratigraphic sections, and unit thicknesses have been determined almost entirely by measurement from maps. The source of each major stratigraphic section is given in table 1. Parts of the sections were measured locally by tape where more detailed descriptions were required for specific purposes.

This report is based almost entirely on field studies, but several hundred thin sections from all parts of the Prichard Formation were examined, and modes were determined for most of them by point counts. However, as noted by Harrison and Grimes (1970, p. 11), the fine grain size of most of the rocks makes microscopic modal analysis extremely difficult. Furthermore, the rocks have been subjected to pervasive low-grade metamorphism, and the texture has undergone significant changes. For these reasons I have not attempted any systematic petrography. Semiquantitative spectrographic analyses were obtained for rocks from the Plains area. Harrison and Grimes (1970) have published analysis of rocks from the more quartzitic phase of the Prichard. I was aided in the summer of 1980 by P.G. DeCelles and in the summer of 1982 by L.J. Saraka. I am indebted to these men not only for their able assistance but also for their friendship.

LITHOLOGY AND METAMORPHISM

METASEDIMENTARY ROCKS

The Prichard Formation, which has been regionally metamorphosed to the greenschist facies, is within the biotite or garnet zone everywhere except in Glacier National Park where the rocks belong to the muscovite-chlorite subfacies. Regional metamorphism was static; the texture is granoblastic, and foliation is uncommon except adjacent to major faults and Mesozoic or Cenozoic intrusions. The rocks are fine grained, drab, hard, and dense. Most workers in Belt terrane classify the rocks of the Belt Supergroup as quartzite, siltite, and argillite; that usage is followed in this report. Inasmuch as the original clay minerals have been

[1]The Prichard Formation has been identified in exposures on the Middle Fork of the Clearwater River between the Atlanta and Bitterroot lobes of the Idaho batholith (Reid and others, 1973), in the Anaconda Range of southwestern Montana (Emmonds and Calkins, 1913, p. 37–39), and in the Highland Mountains south of Butte, Mont. (McMannis, 1963, p. 37–39; Thorson, 1984). However, the exposures on the Middle Fork of the Clearwater are high-grade metamorphics whose assignment to the Belt Supergroup is questionable (J.E. Harrison, oral commun., 1987), and the rocks assigned to the Prichard in the Anaconda Range have more recently been identified as contact-metamorphosed rocks of the Missoula Group (Elliot and others, 1985). The rocks in the Highland Mountains identified by McMannis as belonging to the Prichard do closely resemble much of that formation in lithic character, and they are in about the same stratigraphic position. However, these exposures are 240 km southeast of the nearest exposures definitely identified as Prichard, and they are best treated as part of the Helena embayment sequence.

TABLE 1.—*Sources of data in the stratigraphic sections*

Section No.	Name	Sources of data
1	Siegel Mountain	Thicknesses measured from Cressman (1981); lithologies from Cressman (1985).
1A	Seepay Creek	Thicknesses of sedimentary section same as in section 1. Thicknesses and stratigraphic positions of sills from unpublished map (scale 1:24,000) by R.L. Earhart and R.E. Van Loenen (written commun., 1981) and from reconnaissance.
2	McLaughlin Creek	Thicknesses measured from Cressman (1981); lithologies from Cressman (1985).
3	Eddy Creek	Thicknesses from traverse up Eddy Creek in northeastern part of St. Regis, Mont., 15-minute quadrangle; rocks not examined closely.
4	Prospect Creek	Section constructed from exposures in south half of Cooper Gulch, Idaho-Mont., 15-minute quadrangle; interval from member E to member G measured uphill from exposures of member E in roadcut along Prospect Creek road 600 m north of Glidden Gulch; member G measured in Cox Gulch; base of transition member located on National Forest road to Driveway Peak; thickness of transition member, which was measured along Cooper Gulch road, is approximate.
5	Maple Peak	Member F measured and described in Bear Gulch in east-central part of Burke, Idaho-Mont., 15-minute quadrangle; member G measured and described from roadcuts in Bear Gulch and from natural exposures on Maple Peak north of Bear Gulch; member H measured along ridge that extends east from Maple Peak and described from roadcuts along National Forest road that parallels Idaho-Montana State line at head of Butte Gulch; transition member measured along Idaho-Montana State line at the head of Butte Gulch. Section east of Maple Peak may be faulted, and thickness of member H is approximate.
6	Pine Creek	Thickness of section and stratigraphic position of quartzites from structure section on plate 1 of Hobbs and others (1965); lithology from reconnaissance.
7	Bear Creek	Measured from roadcuts along logging roads and from natural exposures between Deep Creek and Bear Creek in Seven Point Mountain, Mont., 7½-minute quadrangle; contact with Burke Formation from Harrison and others (1986). Member E may crop out in inaccessible exposures on north side of Bear Creek.
8	Stanton Mountain	Lower part generalized from section measured by R.E. Van Loenen (written commun., 1984); upper part generalized from section measured by J.W. Whipple and J.C. Sample. Basal part exposed in McDonald Creek in western part of Glacier National Park; remainder measured on Stanton Mountain, also in western part of Park. I have examined only part of section.
9	Little Wolf Creek	Composite of surface section 1,800 m thick and subsurface section penetrated by Gibbs No. 1 borehole drilled by Atlantic Richfield and Marathon Oil Companies. Surface section based on my mapping. Subsurface section based on geophysical logs and on information on lithology from industry geologists. Subsurface section not corrected for dip which is mostly less than 10° in upper three-fourths and less than 20° in lower one-fourth of section. Subsurface section cut by Pinkham thrust, probably at about base of member E, but the thrust probably repeats very little section. (Boberg, 1985, located the Pinkham thrust at about 17,000 ft, or 5,200 m, and inferred a throw of about 7,000 ft, or 2,100 m.)
10	Callahan Creek	Upper half of section based on my mapping in Troy and Spar Lake, Mont., 7½-minute quadrangles; section from middle down to key bed B from traverse along ridge from Smith Mountain west to Mt. Willard in north part of Mt. Pend Oreille, Idaho-Mont., 15-minute quadrangle; basal part from map by Harrison (1969) and my reconnaissance near Lunch Peak in west-central part of Mt. Pend Oreille quadrangle. Thickness of interval from argillite member to upper member uncertain.
11	Yaak River	Section based on geologic mapping at scale of 1:48,000 (Cressman and Harrison, 1986); position of sills is that on east side of Sylvanite anticline near Yaak River.
11A	Northwest Peak	Placement of sills and key beds based on my mapping in Northwest Peak, Mont., 7½-minute quadrangle (Cressman and Harrison, 1986) and on mapping by Burmester (1985) in Canuck Peak, Idaho-Mont., 7½-minute quadrangle.
12	Bead Lake	Based on descriptions and mapping (scale 1:62,500) by Miller (1974a, b, c, d) supplemented by reconnaissance.

TABLE 1.—*Sources of data in the stratigraphic sections*—Continued

Section No.	Name	Sources of data
13	Chewelah	Section from description and mapping by Miller and Clark (1975, p. 4–6; pls. 1, 2) supplemented by reconnaissance. Position of sills with respect to quartzite-bearing part of section from ridge that extends southeast from Chewelah Mountain (Miller and Clark, 1975, pl. 1); middle sill of Miller and Clark (1975, pl. 1) assumed to be same as upper sill that trends along Cottonwood Divide (Miller and Clark, 1975, pl. 2) because both sills are within quartzite section; thickness from this sill to top of Prichard Formation from Miller and Clark (1975, pl. 2); lower part of section and lowest sill from roadcuts along logging road in sec. 4, T. 31 N., R. 42 E., northeast of Nelson Peak (Miller and Clark, 1975, pl. 2); upper part of section examined in roadcuts in sec. 30, T. 32 N., R. 42 E. Excellent exposures of quartzite-bearing section along Flowery Trail road, sec. 5, T. 32 N., R. 42 E. Prichard Formation in this area is overturned, faulted, and intruded by quartz monzonite of Cretaceous age. Although sequence shown in stratigraphic section is probably correct, thicknesses are approximate. Metamorphic recrystallization of micas and bedding-plane shearing have destroyed sole marks.
14	Katka Face	Section based on mapping of Moyie Springs, Curley Creek, Leonia, and Clefty Mountain, Idaho, 7½-minute quadrangles by R.F. Burmester (1986). Parts of section are well exposed along Katka Face road.
15	Upper Whitefish Lake	Section from exposures in sec. 3, T. 33 N., R. 23 W., and secs. 16, 17, and 21, T. 34 N., R. 33 W., on east side of Upper Whitefish Lake. Whitefish Lake, Mont., 7½-minute quadrangle. Thicknesses from mapping by Whipple and Harrison (1987); lithology from reconnaissance. Member H traversed by fault, and thickness is estimated.
16	Osburn	Section from exposures on north side of valley of South Fork of Coeur d'Alene River between Kellogg and Osburn, Idaho; thicknesses uncertain because of complex structure; thicknesses shown measured from Terror Gulch on plate 3 of Hobbs and others (1965); dip taken as 42°, which is average dip of Prichard Formation in section *B–B'*, Hobbs and others (1965, pl. 2); base of member D approximately at north abutment of bridge over South Fork of Coeur d'Alene River; top of member E about 1,000 m up Terror Gulch.

recrystallized, argillite refers to rocks that consist mostly of phyllosilicates rather than to rocks that consist of clay-size material.

Mesozoic compression and intrusion superposed dynamic metamorphism on the earlier static metamorphism in the westernmost and southernmost parts of the area, and rocks of the Prichard Formation there have been converted to mica schist, biotite-quartz-plagioclase gneiss, and quartzite. These occur in the shaded outcrop areas on plates 1 and 2. Most of the schists are of the greenschist facies, but based on the mapping of isograds by Hietanen (1968, pl. 2) and the identification of formations by Harrison (Harrison and others, 1986), some are of the amphibolite facies and are in the sillimanite-kyanite zone. The Aldridge Formation in Canada has also been metamorphosed mostly to the greenschist facies, though some rocks there attain the amphibolite grade (Monger and Hutchinson, 1971).

The following discussion applies only to the statically metamorphosed rocks that make up the bulk of the Prichard Formation.

The mineralogic compositions of quartzites, siltites, and argillites of the Prichard and Aldridge Formations compiled from various sources are given in table 2. In columns 3 through 6, 8, and 10, which are based on point counts of thin sections, quartz and plagioclase are combined because X-ray diffraction shows that the plagioclase (mostly albite) content is generally underestimated in thin sections (Harrison and Campbell, 1963, p. 1416, table 1). Note that "argillite" as used in column 11 includes both argillite and siltite, so the results in that column are not directly comparable to the other argillites listed. Table 3 gives the chemical composition of rocks from the two formations. Where the analyses are for single samples, the mineralogic compositions determined by J.E. Harrison from X-ray diffraction are also given. Again, "argillite" as listed in column 11 includes both argillite and siltite.

The quartzites of the Prichard and Aldridge Formations, based on their mineralogic composition and texture, would be classified as graywackes by most definitions of that term. Feldspar constitutes one-third to one-fourth of the detrital grains. The matrix, now metamorphosed to sericite, biotite, and chlorite, makes up from 10 to nearly 50 percent by volume of the rock, and the detrital fraction is generally very fine to fine

TABLE 2.—*Mineralogic composition of rocks from the Prichard and Aldridge Formations*

[$\bar{x}$, mean; $\hat{\sigma}$, standard deviation; leaders (--), not determined]

Composition	Quartzite							Siltite		Argillite			
	1	2		3	4	5	6	7	8	9	10	11	
		$\bar{x}$	$\hat{\sigma}$									$\bar{x}$	$\hat{\sigma}$
Quartz	63	60.3	11.32	58	66	61	77	41	54	30	26	35.0	7.30
Plagioclase	21	22.7	8.65					24		19		17.5	7.87
K-feldspar	Trace	0.3	0.54	0	Trace	0	Trace	2	Trace	Trace	0	1.5	4.02
Biotite	2	5.8	5.79	10.5	6	6.5	Trace	9	21	10	22	15.4	9.90
Chlorite	5	0.5	0.76	Trace	Trace	10.5	0	4	2	8	2	0.7	1.74
Muscovite		2.2	1.95	0	4	11.5	23	--	0	--	0	2.6	4.08
Sericite	9	6.9	6.89	31	21	10		20	24	33	47	23.3	13.60
Graphite?	--	0.1	0.35	0	0	0	0	--	Trace	--	Trace	1.5	1.92
Fe-ores	--	0.5	0.83	Trace	Trace	Trace	Trace	--	Trace	--	Trace	1.1	1.28
Sphene	--	0.2	0.37	Trace	Trace	Trace	Trace	--	Trace	--	Trace	0.4	0.74
Carbonate	Trace	0.6	1.65	0	Trace	0	0	0	0	Trace	0	1.1	2.41
Total	100	100.1	--	99.5	97	99.5	100	100	101	100	97	100.1	--

Column: 1. Average of 6 samples from Pend Oreille Lake area, Idaho; determined by X-ray diffraction supplemented by flame photometer analysis for K and Na; Harrison and Campbell (1963, p. 1417).

2. Mean and standard deviation for samples from the Aldridge Formation near Kimberly, British Columbia, Canada; Edmunds (1977a).

3. Average of 4 samples from the Yaak area, Montana; determined by point count.

4. Average of 6 samples from outcrop area west of Moyie thrust; determined by point count; two samples contain 18 and 11 percent albite and 66 and 58 percent quartz.

5. Average of 2 samples from Cooper Gulch, Idaho-Mont., 15-minute quadrangle; determined by point count.

6. Average of 5 samples from member E; determined by point count.

7. Average of 7 samples from Pend Oreille Lake area, Idaho; determined by X-ray diffraction supplemented by flame photometer analysis for K and Na; Harrison and Campbell (1963, p. 1417).

8. Average of 6 samples from Yaak area, Montana; determined by point count.

9. Average of 11 samples from Pend Oreille Lake area, Idaho; determined by X-ray diffraction supplemented by flame photometer analysis for K and Na; Harrison and Campbell (1963, p. 1417).

10. Average of 6 samples from the Yaak area, Montana; determined by point count.

11. Mean and standard deviation for samples from near Kimberly, British Columbia, Canada; Edmunds (1977a); includes siltite.

grained and poorly sorted. In comparing these percentages with those of younger, unmetamorphosed rocks it should be remembered that the rocks of the Prichard and Aldridge are nearly totally consolidated and that some of the phyllosilicates may originally have been framework grains rather than matrix. The quartzites of both formations contain more SiO_2 and less Al_2O_3 than do the graywackes listed by Pettijohn (1963, tables 6–8) and are more similar in that respect to his arkoses and subarkoses. Conversely, the graywackes are similar to those listed by Pettijohn (1963, figs. 2, 3) in that the Na_2O/K_2O ratios of the quartzites are greater than 1, whereas the ratio is less than 1 in the associated argillites.

By Gilbert's 1954 (p. 292, 293) classification the quartzites are feldspathic wackes, but if most of the matrix was orginally deposited as rock fragments, as some observations suggest, the quartzites would have been lithic wackes before diagenesis and metamorphism.

Because of metamorphic recrystallization, both siltites and argillites of the Prichard Formation consist of grains that are mostly of silt size, though mineralogically the argillites consist mostly of phyllosilicates and undoubtedly originated as shales. Most of the argillite occurs in graded siltite-argillite couplets or as laminae and thin beds that alternate with siltite layers.

Table 4 compares the chemical composition of argillites of the Prichard and Aldridge Formations with that of Archean and Early Proterozoic shales (mostly metamorphosed to the lower greenschist facies) from the Canadian shield and of Phanerozoic shales from a wide variety of localities. The Prichard and Aldridge argillites are very similar in composition to the Early Proterozoic shales of the Canadian shield, but they contain more SiO_2, more Na_2O and K_2O, and less combined FeOand Fe_2O_3 than do the Phanerozoic shales. Most of these differences reflect the greater total silt content and the more abundant feldspar of the Prichard and Aldridge. The marked difference in composition between the Archean and the Proterozoic shales probably reflects more mature weathering and increasing differentiation of the source igneous rocks (Cameron and Garrels, 1980, p. 184, 185). The Na_2O/K_2O ratio of the Aldridge argillites is 0.42 and of the Prichard argillites, 0.40; these are greater than that of 0.34 for the

TABLE 3.—*Elemental and mineralogic composition of rocks from the Prichard and Aldridge Formations*

[$\bar{x}$, mean; $\hat{\sigma}$, standard deviation; leaders (--), not determined]

Composition	Quartzite			Siltite				Argillite					
	1	2		3	4	5	6	7	8	9	10	11	
		$\bar{x}$	$\hat{\sigma}$									$\bar{x}$	$\hat{\sigma}$
SiO_2	83.90	78.893	5.738	70.82	79.74	79.59	77.10	65.05	65.62	67.04	64.68	64.162	3.219
Al_2O_3	8.45	9.415	2.826	14.88	11.52	10.64	11.15	18.10	17.95	17.54	18.09	16.055	2.762
Fe_2O_3	.87	--	--	.76	1.26	.60	1.40	2.23	.95	1.63	.99	--	--
FeO	.72	2.435	1.073	3.38	.63	2.07	2.25	2.40	3.77	2.35	3.74	4.606	1.198
MgO	.36	0.778	.458	1.11	.58	.70	1.01	1.35	1.41	1.65	1.53	--	--
CaO	.11	1.359	1.080	.69	.01	.15	.13	.24	.18	.21	.32	1.962	1.817
Na_2O	2.81	2.439	.997	2.30	2.82	2.01	2.33	1.85	2.11	1.26	1.69	1.723	.673
K_2O	1.47	1.476	.667	3.64	1.84	2.33	1.91	4.79	4.20	3.59	4.60	4.136	1.166
H_2O+	.54	--	--	1.45	.89	1.12	1.73	2.87	2.74	3.27	2.73	--	--
H_2O-	.08	--	--	.03	.12	.01	.03	.27	.08	.33	.09	--	--
TiO_2	.38	.409	.136	.56	.35	.36	.38	.69	.70	.61	.72	.681	.168
P_2O_5	.04	--	--	.07	.02	.03	.06	.02	.08	.01	.07	--	--
MnO	.03	--	--	.06	.02	.04	.04	.06	.05	.06	.07	--	--
CO_2	.01	--	--	.02	.01	.02	.06	.01	.02	.01	.14	--	--
SO_3	--	--	--	--	--	--	--	--	--	--	--	--	--
Quartz	61	--	--	46	54	53	41	35	20	39	28	--	--
K-feldspar	--	--	--	--	--	--	--	--	--	--	3	--	--
Plagioclase	26	--	--	22	23	22	19	20	18	12	19	--	--
Biotite	3	--	--	8	6	6	--	5	10	4	8	--	--
Chlorite	--	--	--	--	--	--	20	--	21	20	15	--	--
Sericite	10	--	--	24	10	19	20	38	31	25	27	--	--

Column: 1. Quartzite from member G, Hot Springs, Mont., 7½-minute quadrangle. E.L. Brandt, analyst.
2. Mean and standard deviation for quartzites of the Aldridge Formation, near Kimberly, British Columbia, Canada. Edmunds (1977a).
3. Siltite from member E, Plains, Mont., 15-minute quadrangle. E.L. Brandt, analyst.
4. Siltite from member G, Hot Springs, Mont., 7½-minute quadrangle. E.L. Brandt, analyst.
5. Siltite from quartzite member, Clark Fork, Idaho-Mont., 15-minute quadrangle. George Riddle, analyst.
6. Siltite from quartzite member, Packsaddle Mountain, Idaho, 15-minute quadrangle. George Riddle, analyst.
7. Dark- and light-gray laminated argillite from quartzite member, Packsaddle Mountain, Idaho, 15-minute quadrangle. R.T. Okamura, analyst.
8. Dark- and light-gray laminated argillite from near top of quartzite member, Clark Fork, Idaho-Mont., 15-minute quadrangle. R.T. Okamura, analyst.
9. Dark- and light-gray laminated argillite from member H, Lonepine, Mont., 7½-minute quadrangle. E.L. Brandt, analyst.
10. Dark- and light-gray laminated argillite from upper member, Packsaddle Mountain, Idaho, 15-minute quadrangle. D.F. Powers, analyst.
11. Mean and standard deviation for argillites and siltites of the Aldridge Formation, near Kimberly, British Columbia, Canada. Edmunds (1977a).

TABLE 4.—*Comparison of major-element composition of shales of Archean, Proterozoic, and Phanerozoic ages*

Composition	Archean[1]	Proterozoic			Phanerozoic[1]
		Early[1]	Middle		
	Canadian Shield	Canadian Shield	Prichard Fm[2]	Aldridge Fm[3]	
SiO_2	60.52	65.00	67.79	66.63	63.11
Al_2O_3	17.63	16.22	18.52	16.68	18.80
[4]Fe_2O_3	7.91	6.37	5.01	5.32	7.79
MgO	3.45	2.51	1.50	2.55	3.11
CaO	2.16	0.52	0.25	2.04	0.07
Na_2O	2.62	1.44	1.79	1.79	1.27
K_2O	2.84	4.83	4.44	4.30	3.69
TiO_2	0.76	0.76	0.70	0.71	0.89
Total	97.89	97.65	100.00	100.02	98.73

[1]Cameron and Garrels (1980, table 3); rocks have been metamorphosed to lower greenschist facies (Cameron and Jonasson, 1972, p. 987–988).
[2]Mean of four argillite samples in table 3, recalculated to 100 percent.
[3]Column 11, table 3; argillite and siltite; recalculated to 100 percent.
[4]FeO and Fe_2O_3 as Fe_2O_3.

Phanerozoic shales and 0.30 for the Early Proterozoic shales, which perhaps reflects less intense weathering in the source area of the Prichard.

METAMORPHISM

Common minerals in the Prichard Formation are quartz, albite, oligoclase, chlorite, biotite, and sericite. These are typical of the quartz-albite-epidote-biotite subfacies of the greenschist facies (Fyfe and others, 1958, p. 223). The quartz is present as detrital silt and sand grains, but solution, overgrowth, and suturing have so altered grains that original shapes are seldom apparent. Oligoclase is present in amounts of less than a few percent, and most of the plagioclase consists of untwinned albite that is difficult to distinguish from quartz in thin section and is most readily detected by X-ray diffraction (Harrison and Campbell, 1963, p. 1416, table 1). Microcline and garnet are present in only a few samples. Biotite occurs as randomly oriented anhedral flakes that average about 40 μm in diameter. Biotite is nearly ubiquitous, but it is most common in argillic siltite and silty argillite. The material herein termed sericite has been identified near Pend Oreille Lake by Maxwell and Hower (1967) and near Alberton by Wells (1974) as the 2M polymorph of illite, probably converted during metamorphism from 1M*d* illite such as that present in the upper part of the Belt Supergroup (Maxwell and Hower, 1967). Chlorite, which is seen in section as randomly oriented poikiloblastic laths several millimeters long, is present in many beds. Minor constituents include carbonaceous matter, pyrite, pyrrhotite, magnetite, tourmaline, and sphene. According to Huebschman (1973, p. 691), the carbonaceous matter could not be detected by X-ray diffraction and probably is amorphous.

Both the grain size and the disorientation of the sericite increase downsection. Near Plains, Mont., the sericite flakes near the top of the Prichard Formation average less than 10 μm in diameter and few exceed 20 μm; the sericite exhibits a preferred but far from perfect orientation parallel to the bedding. Near the base of the section many sericite flakes exceed 40 μm in diameter, and their orientation is random (Cressman, 1985, p. 78). In the exposure of the Prichard Formation north and south of the Yaak River north of Troy, Mont., sericite flakes near the top of the section average 10–15 μm in diameter and reach a maximum of 20 μm; at the base the average diameter is 60 μm and the maximum is 120 μm.

The most common mineral assemblage in the quartzite is quartz-albite-sericite-biotite. The most common assemblage in argillite is also quartz-albite-sericite-biotite; quartz-albite-sericite-chlorite-biotite is less common, and a quartz-albite-sericite-chlorite-biotite-garnet assemblage is present in argillite of member D of the Prichard Formation in the Plains-Perma area of Montana (see plate 4 for stratigraphic nomenclature). I have not determined the composition of these garnets, but the relatively high manganese content of member D suggests that the garnet is spessartine, which is stabler at lower temperatures than are other garnets (Hsu, 1968). A quartz-albite-biotite-sericite-calcite assemblage is present locally in the upper part of the Prichard. Microcline, dolomite, quartz, and biotite occur together in part of section 9, but I do not know the complete assemblage.

Widely spaced, nearly white quartzite lenses several centimeters thick and several meters long that contain garnet and chlorite or hornblende and, locally, calcite (J.E. Harrison, written commun., 1987) occur in several parts of the section. The rock consists of sutured silt-sized to very fine sand sized quartz grains, poikiloblastic garnet, and chlorite or hornblende, in part poikiloblastic, in laths several millimeters long. The most conspicuous exposures are in roadcuts through the upper member of the Prichard Formation on the east side of Lake Koocanusa where the sparse white quartzite layers contrast with drab, rusty-weathering argillite of the upper member. These lenses occur in the same stratigraphic position as far west as the exposures immediately west of Yaak, Mont. One thin section of a lens from near Lake Koocanusa consists of quartz, hornblende, garnet, and minor albite. Near Plains similar lenses occur near the base of member E. A thin section from there consists of quartz, chlorite, and garnet. Also near Plains, some slump-fold units in member B are directly underlain by quartzite lenses; a sample from one lens that was examined in thin section consists of quartz, andesine, hornblende, and garnet. Finally, a white quartzite lens at the top of a quartzite turbidite bed from exposures west of Troy and north of Pend Oreille Lake consists of quartz, garnet, epidote, and actinolite. This lens is from about 3,500 m below the top of the Prichard Formation.

Most of the assemblages in the white quartzites suggest the quartz-albite-epidote-almandine subfacies of the greenschist facies (Fyfe and others, 1958, p. 223), though the andesine in one of the samples from near Plains suggests the amphibolite facies. The mineral assemblages in the lenses suggest a slightly higher grade than that of the surrounding rock, and the lenses may have been closed systems. The positions of these assemblages on ACF and AKF diagrams suggest that the protolith was dolomitic siltite.

As noted above, generally random orientation of biotite and chlorite porphyroblasts and the increasingly random orientation of sericite downsection indicate that the regional metamorphism was static. Peale (1893, p. 19) noted early that metamorphosed rocks of the Belt Supergroup are overlain unconformably by the unmetamorphosed Flathead Quartzite, and the regional metamorphism was therefore of Proterozoic age (Harrison, 1986). Metamorphic biotite from the upper 500–1,300 m of the Prichard Formation north of Alberton, Mont., gave a K–Ar age of 1,330±45 Ma (Obradovich and Peterman, 1968). This is the age at which the biotite last cooled below the argon-retention temperature, which is 250–350 °C for a cooling rate of 30 °C/m.y. (Dodson and McClelland-Brown, 1985). Approximately the same part of the section is dated as younger than 1,433±30 Ma by the Crossport C sill (Zartman and others, 1982). Therefore, the age of burial metamorphism for the Prichard is between 1,330±45 Ma and 1,433±30 Ma.

Precambrian metamorphism was accompanied by more severe folding in British Columbia than south of the 49th parallel, and at places a spaced cleavage developed (McMechan and Price, 1982, p. 483). The Hellroaring Creek stock that intruded metamorphosed and folded rocks of the Aldridge Formation (Leech, 1962) has been given a minimum age of 1,305±52 Ma by the Rb-Sr whole-rock isochron method (Ryan and Blenkinsop, 1971; McMechan and Price, 1982, p. 481).

The temperature gradient during metamorphism is inferred in figure 3. The stratigraphic column of the Belt Supergroup is composite; that part from the Mount Shields Formation down is for the area between the Ninemile and St. Marys faults (fig. 32), and the part from the Mount Shields up is for the central part of the Wallace 1°×2° quadrangle. The ranges of the metamorphic minerals are from Wells (1974), Harrison (written commun., 1986), and this paper. The upper limit of metamorphic chlorite in this section is not known, but near Alberton chlorite extends to the top of the Garnet Range Formation (Wells, 1974), which is approximately equivalent to the upper part of the Libby Formation.

The rocks of the Belt Supergroup are now nearly completely consolidated, but at the end of Belt deposition the upper part of the section must have been more porous and thicker than at present. In the Elmore No. 1 borehole in the Salton Sea geothermal field in California, porosity at the top of the biotite zone is about 5 percent, but it increases to 15–20 percent in the chlorite zone (McDowell and Elders, 1980). I have therefore assumed by analogy with the Elmore No. 1 borehole that the Belt section above the biotite zone had a porosity of 20 percent at the end of Belt deposition and that the porosity at the top of the biotite zone was 5 percent at the end of Belt deposition. The present-day porosity at the top of the Libby is taken as 5 percent. The porosity-depth curve down to a depth of 5,000 m is from the COST B–2 well drilled near the edge of the Atlantic continental shelf off Delaware (Watts, 1981, fig. 8); below a depth of 5,000 m, the curve is projected. The porosity-depth curve has been positioned so that 20 percent porosity coincides with the top of the Libby Formation in the decompacted section and 5 percent porosity with the top of the biotite zone. The top of the curve, and thus the top of the Belt Supergroup at the end of deposition, is about 2,000 m above the present top of the section. Thus the original thickness of the Belt is estimated to have been at least 19,000 m.

The temperature-depth curve at the end of Belt time in figure 3 was constructed from temperatures inferred from the ranges of the metamorphic minerals. The thermal gradient given by the curve in the top 1 km is 1×10^{-3} °C/cm (5.5 °F/100 ft). Gradients this high have been reported from the Canadian cordillera but not from the conterminous United States (American Association of Petroleum Geologists and U.S. Geological Survey, 1976). For the entire thickness of 19 km the gradient averages 0.19×10^{-3} °C/cm (1.1 °F/100 ft).

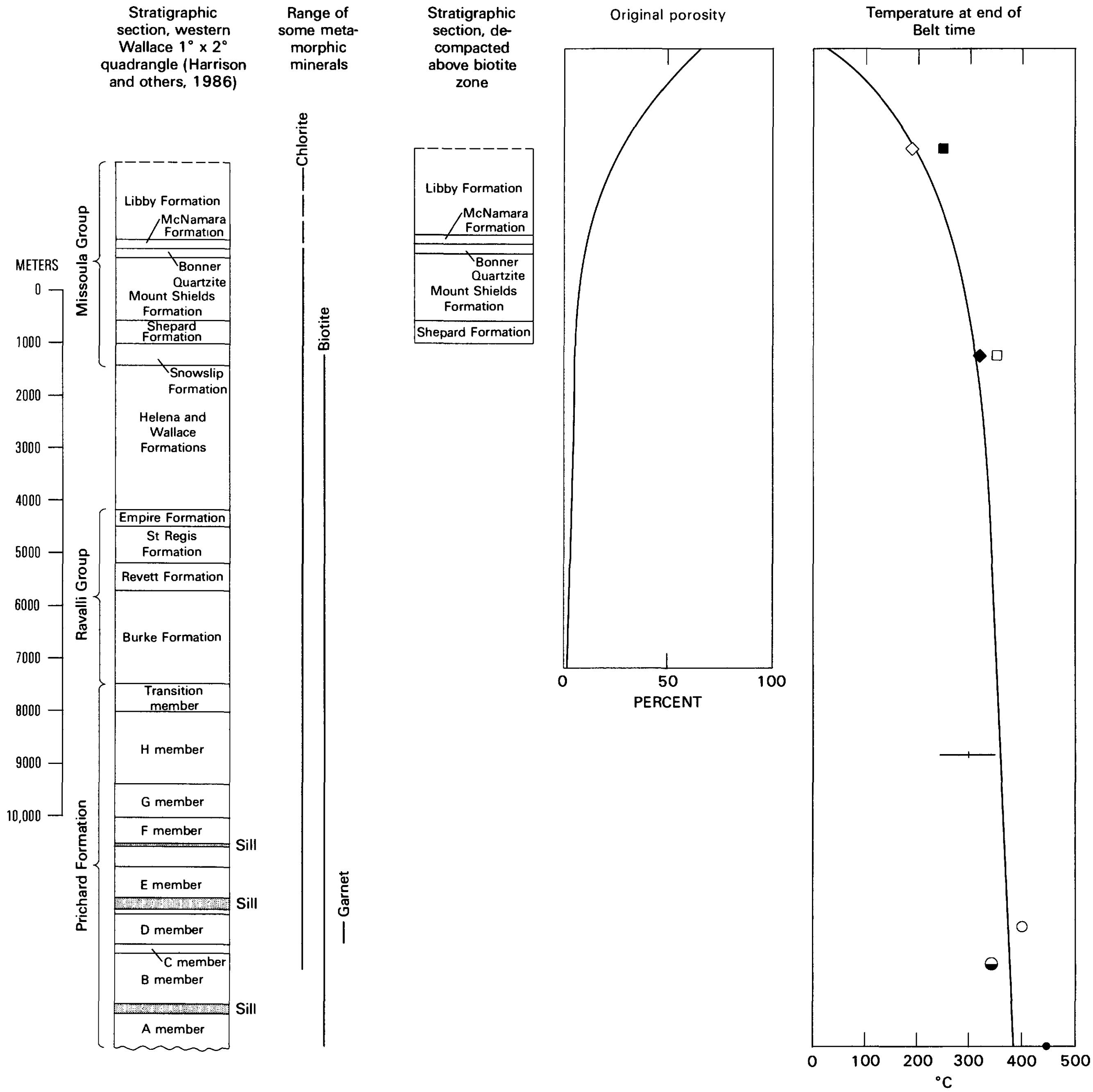
The argon-retention temperature for biotite at the horizon sampled by Obradovich and Peterman (1968) is to the left of the temperature-depth curve. This suggests that the K–Ar date of 1,330±45 Ma for the metamorphic biotite may be slightly younger than the time of metamorphism.

SILLS

Sills of basic to intermediate composition that range in thickness from less than 1 m to as much as 600 m have intruded the Prichard and are common wherever the middle and lower parts of the formation are exposed. I have not studied these rocks, and the following is taken from Bishop's (1973) description of sills that crop out near Crossport, Idaho, on the banks of the Kootenai River opposite the mouth of the Moyie River.

The sills are composed of dark-gray to greenish-black very dense gabbro, diorite, quartz diorite, and granophyre with an igneous texture that is well preserved despite deformation and metamorphism (Bishop, 1973, p. 24). The idealized sill was described as follows by Bishop (1973, p. 32):

> An idealized Crossport sill is composed of a basal chilled gabbro that grades upward with increasing grain size to diabase. Overlying the diabase, again with gradational contact, are medium to coarse grained rocks with subophitic or gabbroic texture that make up the bulk of a sill. The upper chilled gabbro is gradational downward to diabase similar in texture to the lower diabase; however, the contact between the upper diabase and the coarse grained rock may be irregular and sharp. The basal chilled gabbro and diabase may comprise as much as 10 percent of the sill thickness; the central, medium to coarse-grained rock usually gradational from gabbro to quartz diorite may represent as much as 85 to 90 percent of the total thickness, and the upper diabase and chilled gabbro seldom exceeds 5 percent of the total thickness.



EXPLANATION

- ● Maximum temperature of greenschist facies (Verhoogen and others, 1970)
- ◒ Highest temperature at which chlorite is stable in Elmorel borehole, Salton Sea geothermal field, Calif. (McDowell and Elder, 1980)
- ◇ Lowest temperature at which chlorite is stable in Elmorel borehole, Salton Sea geothermal field, Calif. (McDowell and Elder, 1980)
- ◆ Lowest temperature at which biotite is stable in Elmorel borehole, Salton Sea geothermal field, Calif. (McDowell and Elder, 1980)
- ○ Spessartine stable at 3kbar in system Al-Mn-Fe-Si-O-H (Hsu, 1968)
- □ Maximum temperature of biotite isograd (Zen and Thompson, 1974, p. 199)
- ■ Maximum temperature of beginning of chlorite zone (Zen and Thompson, 1974, p. 199)
- —+— Mean and range of Ar-retentium temperature in biotite (Dodson and McClelland-Brown, 1985) shown at horizon from which metamorphic biotite gave K-Ar age of 1330±45 Ma (Obradovich and Peterman, 1968; stratigraphic position from J. E. Harrison, written commun., 1986)

FIGURE 3.—Inferred geothermal gradient at end of deposition of Belt Supergroup.

Major minerals in the sills are calcium amphiboles of the hastingsite group, plagioclase, quartz, biotite, and ilmenite; the range of primary plagioclase composition is from An_{62} to An_{12} (Bishop, 1973, p. 35, 40). Any magnetite originally present was destroyed by autoalteration, and the sills are nonmagnetic.

According to Bishop (1973, p. 41) chemical analyses indicate that the original magma was basaltic; normative quartz and hypersthene characterize the basalt as oversaturated tholeiite, and the rock has a low alkali content characteristic of continental tholeiites.

STRATIGRAPHY

NOMENCLATURE

This report follows Calkins (1909) and all subsequent reports in applying the name Prichard Formation to all metasedimentary rocks of the Belt Supergroup within the report area that lie below the Ravalli Group. As so used, the formation comprises the two major facies noted by Calkins—the section dominated by banded argillite that is present in the type area and extends from Perma, Mont., west to Lake Coeur d'Alene and the section containing abundant gray quartzite that extends from Pend Oreille Lake north to Canada and from Yaak, Mont., in the bend of the Kootenai River, west to Chewelah, Wash. Rocks of each facies were divided into informal members during geologic mapping.

The dominantly argillite section between Plains and Perma, Mont., was divided into eight members designated A through H, all of which are mappable at a scale of 1:62,500 (Cressman, 1981, 1985) (pl. 4, secs. 1 and 2). In this report, about 550 m of beds in the Plains area that overlie member H and that were previously included in the overlying Burke Formation (Cressman, 1985, p. 37–42) are placed in the Prichard Formation as the transition member to make the contact of the Prichard and Burke Formations accord with its placement in the type area of the Prichard. This nomenclature of nine members, developed in the Plains-Perma area, is applicable throughout the area of the argillite facies, though not all members are present throughout the entire extent.

Exposures of the quartzite-bearing facies both north and south of the Yaak River were divided during mapping at a scale of 1:48,000 (Cressman and Harrison, 1986) (pl. 4, sec. 11) into a quartzite member, a massive member, an argillite member, an upper member, and the transition member. The quartzite member, which makes up most of the formation, consists of about 65 percent quartzite. The massive member is intercalated in the lower part of the quartzite member, and the argillite member is intercalated in the upper part of the quartzite member. The quartzite member is overlain by argillite of the upper member which is in turn overlain by the transition member. The massive member is of local extent, but otherwise this terminology can be applied throughout the quartzite-bearing facies.

The above nomenclature, developed by the exigencies of geologic mapping, is used throughout this report. The nomenclature is clumsy and is not in accord with the North American Stratigraphic Code in that the Prichard Formation includes sections of considerably different aspect. A glance at the lines of section (pl. 4) will suggest more appropriate nomenclature to many readers. Nevertheless, I have not attempted to change the nomenclature because this is a reconnaissance study and because any significant revision should be made in consultation with other geologists interested in these rocks and in concert with our Canadian colleagues.

The sills within the Prichard Formation were referred to as the Moyie sills by Kirkham and Ellis (1926) and by Erdmann (1941, p. 11). However, the name is inappropriate in that Daly (1905), who coined the term, applied it to a thick sill (or group of adjacent sills; Daly, 1912, p. 22) that crops out on Moyie Mountain (Border Mountain) on the international border just west of Eastport, Mont., and not to all sills that intrude the lower part of the Belt. Schroeder (1952, p. 21, 22) named the sills in the outcrop area northeast of Newport, Wash., the "Marshall diorite", but the name was not used by Miller (1974a, b, c, d) who remapped the area nor by anyone else. Bishop (1973) applied the name Purcell, introduced by Schofield (1915, p. 56) for all sills in the Purcell Supergroup, to sills on the United States side of the border. However, most workers south of the 49th parallel have followed Calkins' (1909) lead in not applying any name, and that practice is followed in this report.

ARGILLITE FACIES

MEMBERS A AND B

Members A and B of the Prichard Formation were mapped and described in the outcrop area between Plains and Perma, Mont. (Cressman, 1981; 1985, p. 9–18) (pl. 1), where they crop out in the cores of several faulted anticlines. Both members consist mostly of closely spaced, graded couplets from less than one to several centimeters thick of siltite and argillite; siltite is dominant. Member B differs from member A in that it contains abundant syndepositional slump folds, and the contact between the two members was placed at a diorite sill below which slump folds are uncommon. In this paper members A and B are discussed together. Their areas of outcrop are shown on plate 2.

By far the most extensive exposures of members A and B are on both sides of Clark Fork and the Flathead River, southeast of Plains, Mont. The thickest sections are at the confluence of the Flathead with Clark Fork and along the lower part of Seepay Creek, which flows into the Flathead from the south about 12 km east of the junction with Clark Fork. The two members as exposed there total about 1,600 m thick, exclusive of the sill that divides the members. The lowest beds of member A differ little from those high in member B, and there is no lithologic change in the lowest exposures to suggest that the base of the unit is near.

Member B is also exposed in the Coeur d'Alene mining district on the north side of the South Fork of the Coeur d'Alene River a short distance east of Kellogg (pl. 2). The exposures extend from the Mines Monument (at the intersection of the Big Creek road with old U.S. Highway 10) about 2.6 km southeast to near the mouth of Terror Gulch. The top of the member is well exposed in the bank of the South Fork of the Coeur d'Alene River 55 m downstream from the bridge over the South Fork on the Terror Gulch road. The maximum thickness exposed is probably about 300 m.

Members A and B are also present in the Gibbs No. 1 borehole (pl. 1, sec. 9) where nearly 1,000 m, exclusive of the diorite sills, were penetrated. Much of the rock is schistose.

The graded siltite-argillite couplets that make up nearly all of the two members (fig. 4) range from a few millimeters to 10 cm in thickness but average 2–3 cm. The siltite, which is mostly argillic and dark gray, forms the basal two-thirds to three-fourths of the couplets. The dark color results mostly from abundant metamorphic biotite. Much of the siltite is itself laminated, and the laminae consist of siltite-argillite couplets, mostly graded, that are 0.5–3 mm thick. These laminae are mostly planar and parallel to bedding, but some are inclined to bedding. Some siltite layers channel a few millimeters into argillite of the underlying couplet. The siltite grades up through a narrow zone into light-gray argillite.

Some siltite layers are well sorted, light colored, and cross-laminated, and their basal contacts channel several millimeters into the underlying bed. These well-sorted siltites tend to occur in zones.

The graded siltite-argillite couplets are probably fine-grained turbidite deposits. Stow and Shanmugam (1980) described an ideal or complete fine-grained turbidite unit as consisting of 9 units designated T_0–T_8 from base to top. Unit T_0 consists of sand or silt exhibiting fading ripples, micro-cross- and parallel lamination, and a scoured load-cast base; T_1 consists of convolute silt and clay laminae; T_2 consists of thin, irregular laminae and low-amplitude climbing ripples; T_3 consists of thin,

FIGURE 4.—Graded siltite-argillite couplets of member B of Prichard Formation.

regular laminae; T_4 consists of indistinct laminae; T_5 consists of wispy convolute laminae; T_6 consists of graded mud; T_7 consists of ungraded mud; and T_8 consists of microbioturbated mud. In terms of these structural divisions most of the couplets of members A and B of the Prichard consist of divisions T_3–T_7; that is, they range from planar-laminated silt at the base to ungraded mud at the top. Those couplets with cross-lamination in the lower part consist of divisions T_0–, T_1–, or T_2–T_7. Division T_8, microbioturbated mud, is not present because there were no metazoans in Middle Proterozoic time, and divisions T_4–T_6 are not obvious because of changes in texture caused by metamorphism.

An interval consisting mostly of quartzite is present several hundred meters below the top of member B in the exposures east of Plains. This quartzite comprises units 41–44 in the measured section described below as Measured Section 1. The quartzite interval is locally as thick as 30 m, but elsewhere, particularly south of the St. Marys fault, it consists of only a few thin quartzite beds. In two localities the quartzite is in channels several meters thick and several tens of meters wide (Cressman, 1985, p. 17). From map measurements the top of the quartzite is as much as 500 m and as little as 250 m below the top of member B. The apparent range in stratigraphic position of the quartzite may result from unmapped structure or from several separate intervals of quartzite, each with limited extent and located at a different stratigraphic level within the member.

The quartzite beds themselves are mostly structureless, but some are parallel laminated, and most grade to argillite in their top few centimeters. There is no evidence, such as crossbedding, of movement by traction, and the quartzites, like the siltite-argillite couplets, were probably deposited from turbidity currents.

Member B in the Plains area contains abundant syndepositional slumps that are either nearly recumbent individual folds mostly 0.3–0.5 m thick or composite sheets as much as 30 m thick of stacked slump folds. These slumps, which make up as much as 15 percent of some parts of member B, are described and illustrated in Cressman (1985, p. 12–16, figs. 5–9).

The slump folds are most abundant below the quartzite unit described above and below the quartzites of member C. A section of the quartzite unit and the beds below, measured in a roadcut on the south side of the Clark Fork 6.5 km southeast of Plains, describes the distribution of the slump folds in that part of member B. The section is described below as Measured Section 1 and illustrated graphically in figure 5. Beds 41–44 are the quartzite unit.

Measured Section 1

Plains 15-minute Quadrangle. Roadcut on River Road on south side of Clark Fork 6.5 km southeast of Plains, Mont. Measured by E.R. Cressman.

	Thickness (meters)
Prichard Formation (part):	
Member B (part):	
47. Diorite sill; not measured	
46. Argillite and siltite; contains a few beds of quartzite; poorly exposed	17.0
45. Argillite	1.4
44. Quartzite; grades up through 0.2 m into overlying argillite	0.8
43. Argillite	0.8
42. Quartzite, medium-gray, blocky-weathering; prominent partings at intervals of 0.5–1 m; some beds contain planar laminations	15.6
41. Interbedded quartzite and argillite: Quartzite is medium gray, in beds and sets 0.3–0.9 m thick; planar laminated in part; beds grade upward through 2–5 cm to argillite. Argillite is in beds 0.04–0.3 m thick	4.8
40. Covered	3.0
39. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; some beds in lower half are truncated at low angles and may be syndepositional slumps	5.5
38. Siltite, dark-gray, blocky-weathering; planar laminated in part; grades to argillite at base and top	1.7
37. Covered	8.7
36. Interlaminated and very thinly interbedded siltite and argillite; poorly exposed	1.2
35. Composite slump sheet: Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; unit consists of several intervals of contorted bedding that alternate with planar-bedded intervals	4.0
34. Slump fold: Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; bedding contorted and at low angle to bedding below	1.2
33. Very thinly interbedded siltite and argillite, in graded couplets 1.5–2 cm thick	4.8
32. Siltite in planar-laminated beds 4–15 cm thick separated by argillite partings	2.0

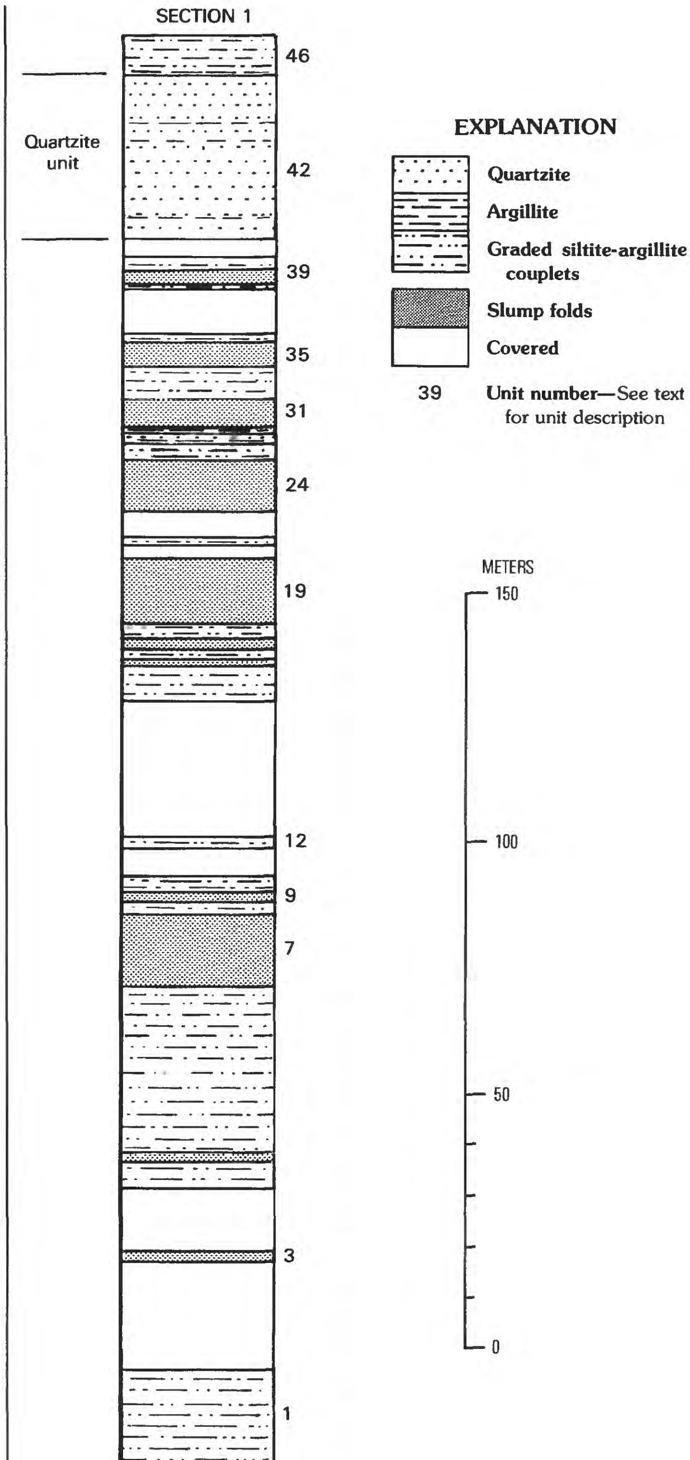


FIGURE 5.—Graphic section of section 1, part of member B of Prichard Formation. Measured in roadcuts on River road 6.5 km southeast of Plains, Mont.

	Thickness (meters)
Prichard Formation (part):—Continued	
Member B (part):—Continued	
31. Composite slump sheet: Siltite in irregular beds 0.3 m thick separated by argillite partings; siltite contains some contorted laminae	4.9

Prichard Formation (part):—Continued	Thickness (meters)
Member B (part):—Continued	
30. Covered	1.2
29. Quartzite; lower half is single bed; upper half consists of several beds, each of which grades upward to argillite	0.9
28. Argillite, well-cleaved	0.1
27. Quartzite, medium-gray, in single bed	0.8
26. Argillite, well-cleaved	1.2
25. Very thinly interlaminated and interbedded siltite and argillite, in graded couplets 1-3 cm thick; siltite is planar laminated	2.5
24. Composite slump sheet: Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; unit consists of several stacked recumbent slump folds	10.3
23. Covered	5.5
22. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets	2.0
21. Covered	1.6
20. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets	0.4
19. Composite slump sheet: Similar to unit above, but consists of stacked recumbent slump folds	13.0
18. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; bed 1 cm thick that consists of quartz, andesine, hornblende, and garnet is 0.3 m below top	2.7
17. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets; bedding contorted in upper 1 m; low-angle discordances between beds in lower 3 m; unit is probably a syndepositional slump	4.0
16. Very thinly interbedded siltite and argillite, in graded couplets 3-4 cm thick; siltite laminated in part	0.7
15. Slump fold: Siltite and argillite similar to unit above	1.0
14. Similar to unit 16; contains some sulfide laminae	5.9
13. Covered	27.0
12. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets	1.5
11. Covered	5.8
10. Interlaminated and very thinly interbedded siltite and argillite, mostly in graded couplets	2.5
9. Composite slump sheet: Interlaminated and very thinly interbedded siltite and argillite, mostly as graded couplets; unit consists of two slump folds	2.0
8. Interlaminated and very thinly interbedded siltite and argillite, in graded couplets 1-2 cm thick; siltite, which is dominant, is laminated	3.1
7. Composite slump sheet: Argillite, silty, medium-dark-gray, hard; bedding obscure, but locally exhibits both contorted and planar laminae	14.0
6. Interlaminated and very thinly interbedded siltite and argillite, in graded couplets 2-8 cm thick; siltite is medium gray, argillite is medium light gray; a few siltite layers channel argillite below, and a few are cross-laminated	28.0
5. Similar to unit above, but contains slump fold 1.6 m thick, 5 m above base	12.0
4. Inaccessible or covered	13.0
3. Slump fold: Argillite, silty, medium-gray; forms single bed; locally contains both contorted and planar laminae	2.8
2. Inaccessible or covered	21.0
1. Interlaminated and very thinly interbedded siltite and argillite, in graded couplets 2-8 cm thick; siltite is medium gray and very locally cross-laminated and locally channels underlying argillite; argillite is medium light gray	19.4

In Measured Section 1 slump folds are most abundant in an interval of about 150 m below the quartzite unit. Similarly, slump folds are more common in the uppermost 100-150 m of member B, immediately below quartzites of member C, than they are in the rest of the member. This suggests that the quartzites were deposited in channels on the submarine slope that grew headward by slumping until they tapped a reservoir of shelf sand (Farre and others, 1983).

The direction of slumping can be inferred by determining the attitude of the axial surface of the slump fold when the adjacent unslumped beds are rotated to horizontal. The direction of slumping is opposite to the dip direction of the rotated axial surface (Woodcock, 1976, p. 171, 172). Slumps are commonly arcuate, prograding slopes may be irregular, and the angle between the axial surface of the slump fold and the adjacent bedding is commonly small and difficult to measure accurately. Therefore, many folds must be measured for the results to have any regional significance.

Forty-seven slump folds were measured in the exposures southeast of Plains. Of these, 24 are north of the St. Marys fault and 23 south of the fault. On plate 2 the poles of the axial surfaces after rotation of bedding to horizontal are plotted on the lower hemispheres of stereographic projections. Statistical computations after the method of Fisher (1953) as elaborated by Steinmetz (1962) and Lindsey (1972, p. 31, 32) give the following mean pole directions: (1) North of St. Marys fault; azimuth = 123°, dip = 78°; and (2) south of St. Marys fault; azimuth = 85°, dip = 84°. These azimuths, which are the inferred direction of slumping, differ by 38°. J.E. Harrison (oral commun., 1985) considers that fold axes south of the St. Marys fault have been rotated counterclockwise about 40°. This would explain the differences in the inferred slump directions. However, the radius of the spherical circle of confidence, P = 0.95, is 8.5° for the pole north of the fault and 10° for that south of the fault; the two circles have considerable overlap, and the difference between the two mean slump directions is not significant.

Angular unconformities of a few degrees may be seen in a few excellent exposures of both members A and B. These may have originated as surfaces stripped by slumping.

The rocks of member B exposed 7 km east-southeast of Kellogg, Idaho (pl. 2), contain slump folds that are smaller and less abundant than those near Plains. Several slump folds in the upper part of the member immediately behind the Miners Monument are about 0.3 m thick; they make up only a small part of the total exposure. I was unable to measure enough axial planes in these exposures to determine the direction of slumping.

The origin of the slump folds is discussed in Cressman (1985, p. 52–54) where it is concluded that they probably formed on the upper part of the depositional slope. The triggering mechanisms could have been oversteepening by progradation, loading by earthquake waves or by storm waves, or reduction of sediment shear strength through rapid loading or through the production of gases during diagenesis. Some slump folds are underlain by a thin layer of quartzite that contains garnet and hornblende; for example, note bed 18 in Measured Section 1. Bed 18 probably was calcareous or dolomitic before metamorphism. Anaerobic decay of organic matter may have produced the carbonate; continued decay may have resulted in the production of methane, and these gases could have reduced the shear strength sufficiently to initiate slumping.

Siltite and argillite of both members A and B have been tourmalinized at six localities in the Plains area. Of these localities, all but one are within 1 km of the St. Marys fault; the sixth is 2.2 km distant. The tourmaline occurs as a felted mass of acicular crystals that can be distinguished with difficulty with a hand lens. The tourmaline, which is black, imparts a sooty appearance to the rock. In some occurrences tourmaline is concentrated in layers that were originally the more argillaceous parts of couplets. Some tourmalinized rocks have also been silicified, and some contain a dusting of sulfides.

MEMBER C

Member C consists of about 100 m of quartzite, interbedded with subordinate siltite and argillite, that directly overlies member B. The member was identified in the exposures between Plains and Perma, Mont. (Cressman, 1981; 1985, p. 18, 19), and surface exposures are restricted to the Plains-Perma area. The areas of outcrop are shown on plate 2. In the Gibbs No. 1 borehole an interval 100 m thick and 4,157 m below the top of the hole is identified as member C on plate 4, but inasmuch as that part of the section is faulted and information on the lithology is scanty, the identification is speculative. Near Kellogg, Idaho, the only other area where the appropriate part of the section is exposed, member C is absent, and member D rests in sedimentary contact on member B.

In its major outcrop area near Plains member C is 75–100 m thick. In one small area (pl. 2, loc. C) member C is missing, and argillite of member D is separated from argillite and siltite of member B only by matrix-supported argillite-pebble breccia a meter or so thick. The breccia formed before regional metamorphism, and member C here probably slumped downslope shortly after deposition.

The character of member C is illustrated by figures 6 and 7. The section (fig. 6) is of approximately the lower half of the member. The quartzite, which is the dominant rock type, is argillic, very fine to fine grained, and medium gray to medium light gray, and weathers to angular blocks. Beds are mostly 0.4–0.6 m thick and are in sets that are 1–10 m thick. Most quartzite beds grade to siltite or argillite in their upper 2–3 cm. Most are structureless, but some are planar laminated (fig. 8). A few sequences in which beds thin upward are present, but for the most part bed thicknesses are arranged randomly.

Sole marks are present but not abundant. At John Tunnel (pl. 2, loc. A) small flute casts and ridge-and-furrow structures are present and indicate a consistent transport direction to the south-southeast. By contrast, at locality B, an exposure along Montana Highway 382, ripple marks are the most common feature, though a few flute casts were found in float. The ripple marks are present as casts at the base of quartzite beds; some are straight-crested and symmetric, whereas others are somewhat sinuous and asymmetric. The straight-crested ripples have wave lengths of 3.5–4 cm and an amplitude of 2 mm. (Assuming an original porosity of 55 percent, the original amplitude would have been about 3.6 mm.) The ripple crests at this locality do not have a preferred orientation (pl. 2), but the asymmetric ripples indicate transport to the south and southwest.

Gradational tops, parallel laminations, and sole marks all indicate that the quartzite beds were deposited from turbidity currents. The straight-crested symmetrical (oscillation) ripple marks at locality B may have resulted from the reworking of the tops of previously deposited turbidites by long-period oceanic waves.

The siltites and argillites of member C, which differ little from those of members A and B, were probably also deposited mostly by low-density turbidity currents.

MEMBER D

In the outcrops between Plains and Perma, Mont., a unit consisting mostly of planar-bedded olive-gray argillite and siltite between the quartzite of member C below and quartzite and irregularly laminated siltite and argillite of member E above was mapped and described as member D of the Prichard Formation (Cressman,

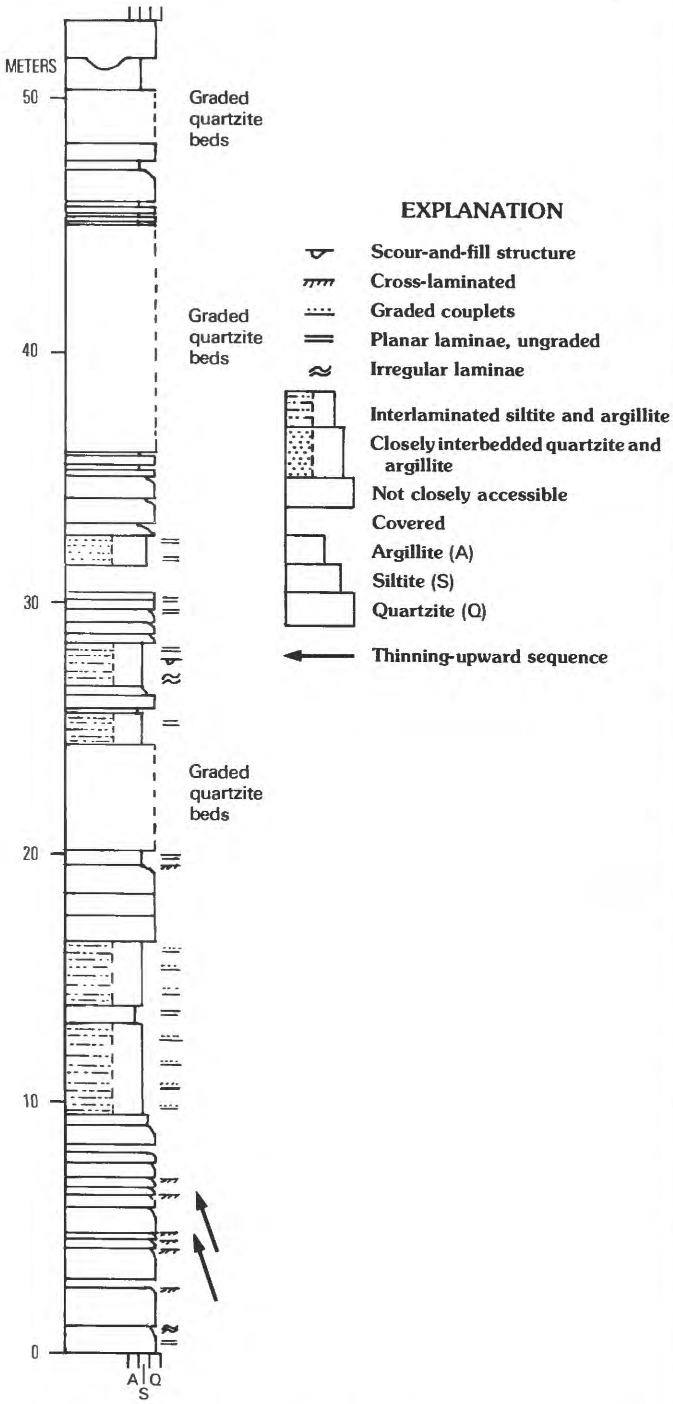


FIGURE 6.—Graphic section of part of member C of Prichard Formation. Measured by P.G. De Celles in cuts along Burlington-Northern Railroad immediately south of John Tunnel, locality A, plate 2.

1981). The distribution of member D in the Plains area is similar to that of member C. Member D is also exposed on the north side of the South Fork of the Coeur d'Alene River between Kellogg and Osburn, Idaho. In the Gibbs No. 1 borehole argillite from 4,011 m to

FIGURE 7.—Member C of Prichard Formation exposed in cut along Burlington-Northern Railroad at south end of John Tunnel, locality A, plate 2. Q, quartzite; A, argillite and siltite.

4,157 m below the top of the hole has been identified as member D, though the identification is as tenuous as that of member C in the same hole.

Member D southeast of Plains is described and pictured in Cressman (1985, p. 19–21), and the following description is from that publication:

> Member D consists of a variety of rock types, but platy-weathering olive-gray silty argillite containing sparse garnet and randomly oriented chlorite porphyroblasts is characteristic. Similar beds are present at least locally in the lower part of member E, but any occurrence of the olive-gray argillite in thicknesses of 50 m or more is diagnostic of member D. The best estimate for the thickness of the member is 290 m, but map measurements indicate some thicknesses as great as 410 m. The greater thickness probably results either from undetected faulting or from inconsistencies in mapping the upper contact.
>
> The basal 70 or 80 m of member D consists of planar-laminated siltite and argillite interbedded with quartzite. The quartzite, which is dark to light gray and somewhat argillic, is in sets mostly 3–6 m thick. Both planar laminae and obscure lenticular bedding may be present. The siltite and argillite are in couplets and are similar to the siltite

FIGURE 8.—Quartzite bed in member C of Prichard Formation. Base of bed is sharp, and top is channeled several centimeters by bed above. Quartzite grades at top of bed through siltite to argillite. Note planar laminae.

and argillite of members A and B. A few small slump folds are present locally.

The overlying three-fourths of the member comprises a coarsening-upward sequence beginning at the base with the platy-weathering olive-gray silty argillite. The silty argillite contains inconspicuous planar siltite laminae. Up section the siltite laminae thicken, and some form lenticular beds several centimeters thick. The lenses are cross-laminated, and the lenticular layers probably originated as a ripple-bedded sheet of silt. Farther up section the siltite layers become still thicker and more closely spaced until they make up most of the rock. The siltite layers, several centimeters thick, commonly have sharp bases and gradational tops. Load casts are present at the base of some. These predominantly siltite beds, in turn, pass upward into siltite with scour-and-fill structure that is commonly interbedded with quartzite.

The upper contact of member D is placed where irregular to wavy laminae and scour-and-fill structure typical of member E become dominant over the planar lamination and bedding typical of member D. At some localities the contact is fairly abrupt and can be located within a few meters. Elsewhere, intervals typical of member D alternate with intervals typical of member E. I do not know whether these two situations represent real differences in the section from place to place or whether they are a function of the type and amount of exposure.

Rocks identified as member D are exposed in the Coeur d'Alene mining district east of Kellogg where they rest directly on the siltite-argillite couplets of member B. The member, which is faulted and not well-exposed, consists of olive-gray siltite and argillite in planar-laminated beds 0.5–10 cm thick. The unit is mostly argillite near the base and is less than 100 m and possibly less than 50 m thick. I did not observe either chlorite porphyroblasts or garnet.

The rocks of member D differ from those of the other siltite-argillite members of the Prichard in that the argillites of member D contain less carbonaceous matter and less iron sulfide, thus accounting for the olive-gray color typical of the member. Furthermore, graded siltite-argillite couplets are not common in the platy-weathering argillite; this suggests that much of member D was deposited directly from suspension. The graded siltite-argillite layers near the top of the member may have formed following storms in the manner suggested by Reineck and Singh (1972) for graded rhythmites on muddy shelves.

MEMBER E

The slabby-weathering argillite and even-bedded siltite and argillite of member D are overlain by nearly 1,000 m of interlaminated siltite-argillite and interbedded well-sorted quartzite that exhibit features such as crossbedding, hummocky cross-stratification, and scour-and-fill structure that are indicative of deposition in shallow water. Mud-chip breccias and desiccation cracks, indicative of periodic subaerial exposure of sediments, are also present in the upper part. These rocks make up member E of the Prichard Formation.

The distribution and thickness of member E are shown on plate 2. The member was first described in the exposures near Plains (Cressman, 1981) where it is about 825 m thick. It has since been identified in the following areas: (1) the Gibbs No. 1 borehole where 950 m of section (exclusive of sills) contains interbeds of well-sorted quartzite typical of member E; (2) near the mouth of Daisy Creek 16 km southeast of Thompson Falls, Mont., where 650 m are present, but the base is not exposed; (3) the north side of Glidden Gulch 14 km east-southeast of Murray, Idaho, where less than 10 m of the uppermost part of the member crops out; (4) east of Kellogg, Idaho, where the member is at least 620 km thick; (5) in the drainage of Pine Creek, Idaho, where the member may be more than 920 m thick; and (6) in roadcuts on U.S. Interstate 90 about 14 km east of the town of Coeur d'Alene, Idaho, where the uppermost part of the member is exposed. Member E may also be present in a fault-bounded block west of Murray, Idaho, and along Bear Creek 24 km north-northwest of Thompson Falls.

Member E in the exposures near Plains was described as follows (Cressman, 1985, p. 22–27):

> Most of member E consists of interlaminated light-gray siltite and dark-gray argillite that commonly occurs as graded couplets. The laminae are mostly less than 1 mm thick, but particularly in the lower part of the member bedding sets several centimeters thick of the very thinly laminated siltite-argillite may alternate with wavy to lenticular beds of siltite several centimeters thick. In general, though, the lamination is considerably thinner in member E than in the other members of the Prichard. The laminae are mostly wavy to crinkly, and planar laminae that are so typical of the rest of the Prichard Formation are uncommon.
>
> The lower part of the member contains scour-and-fill structures 15–20 cm deep that are filled by interlaminated siltite and argillite that drapes into the scour. Many of the siltite layers in the lower part are irregular in thickness, and some are lenticular. The combination

of scour and fill, irregular and lenticular beds, and the effects of differential compaction impart an irregular wavy appearance to this part of the section. Higher in the member the thicker layers become less common, the laminae become irregular to crinkly, and mud cracks and mud-chip breccias become common. The mud cracks are well formed and polygonal.

Quartzite forms a minor but locally conspicuous part of member E. The quartzite, considerably less sericitic than quartzite in member C, is in beds mostly about 0.5 m thick and in sets as thick as 9 m. Flaser bedding, ripple bedding, and crossbedding, some of the herringbone type, are common. Planar and trough crossbedding have been observed in sets as thick as 0.7 m. Wedge-shaped sets of high-angle, small-scale planar crossbeds are present in one quartzite bed that is well exposed in the first gully north of the mouth of Siegel Creek. Most of the quartzite is fairly clean, but the dark laminae are sericitic. Quartzite beds are present throughout the member but are most numerous in the lower one-third. Individual quartzite units do not seem to be continous. Planar-laminated quartzites, such as those in member C, are not present.

Platy-weathering olive-gray silty argillite similar to that which characterizes member D is present at least locally in the lower part of member E where it constitutes an interval a few tens of meters thick. These occurrences suggest that members D and E intertongue. On the mountainside east of Quinns Hot Springs and on the west side of Seepay Creek, this interval contains a thin, speckled, nearly white bed that consists of quartz, chlorite, and garnet.

The upper contact of member E is sharp and marked by the abrupt upward passage from the thinly laminated siltite-argillite exhibiting various shallow-water structures to planar laminated and very thinly interbedded argillite and silty argillite of the overlying member F.

Some of the features described above as a combination of scour-and-fill structure and lenticular bedding are actually hummocky cross-stratification. One such feature is illustrated by figure 12 of Cressman (1985).

The exposure at Daisy Creek is too fragmentary and that at Glidden Gulch too small for detailed description, but both contain the blocky-weathering wavy-laminated siltite and argillite that are characteristic of member E.

East of Kellogg, Idaho, member E is exposed on both the north and south sides of the South Fork of the Coeur d'Alene River. South of the river the member is truncated by the Osburn fault, but both the upper and lower contacts are present in the outcrop belt north of the river. However, the member thins westward in these northern exposures because of internal faults, and the thickness of 620 m measured from the map must be a minimum. The basal beds of the member east of Kellogg consist of very thinly interbedded siltite and argillite, some of which form graded couplets that are as much as 12 cm thick. The bedding is wavy, and some siltite and some couplets fill scours as deep as 12 cm (fig. 9*A*). Some of the thicker graded couplets that exhibit abundant flute casts were probably deposited from low-density turbidity currents that flowed over very irregular surfaces (fig. 9*B*). Higher in the section but still in the lower part of member E the siltite and argillite layers become thinner, and siltite becomes dominant. The siltite is in laminae and beds 0.1–20 mm thick that make up about 75 percent of that part of the section.

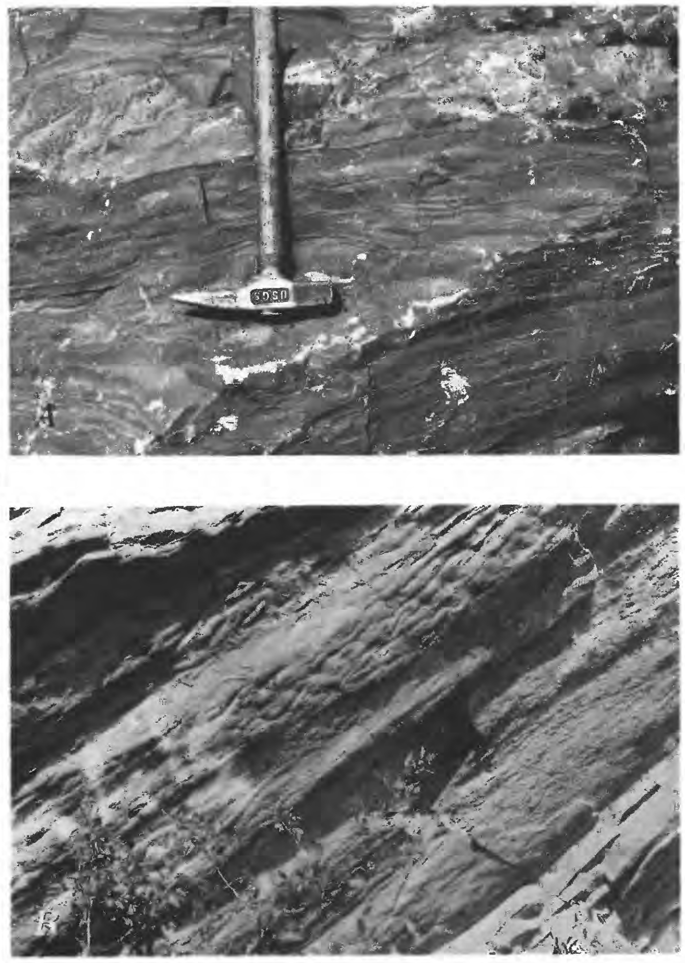


FIGURE 9.—Photographs of beds near base of member E of Prichard Formation east of Kellogg, Idaho. *A*, Graded siltite-argillite couplets. Siltite, light gray; argillite, dark gray. Note that many siltite layers are lenticular. *B*, Flute casts. Stratigraphic position about same as *A*.

The argillite is in laminae from less than 1 to 10 mm thick. A few graded couplets are present. The bedding is wavy, some siltites are lenticular, some couplets fill scours as much as 10 mm deep, and some beds are cross-laminated. The major part of the member consists of medium-gray siltite and very fine grained quartzite in beds 5–8 cm thick and in sets as much as 3 m thick that alternate with sets 1.8–2.5 m thick of interlaminated siltite and argillite similar to that described in the preceding paragraph. Siltite and quartzite sets increase in thickness and abundance upsection to make up about half of the upper part of the member. Many of the siltite and very fine grained quartzite beds are lenticular, and some of the sets exhibit hummocky cross-stratification. The upper part contains at least five beds or sets of light-gray well-sorted quartzite (Hobbs and others, 1965, p. 32). The thickest of these thins eastward from about 80 m to no more than 20 m in a distance of 3 km

(Hobbs and others, 1965, pls. 2, 3). The contact, which is sharp, of member E with planar-laminated argillite and siltite of member F is well exposed on the north side of old U.S. Highway 10 about 533 m west-northwest of the entrance to the Montgomery Gulch road (see Hobbs and others, 1965, pl. 2 for location).

Member E is present on both sides of Pine Creek southwest of Kellogg. Measurements from plate 1 of Hobbs and others (1965) give a thickness of 920 m; the base is not exposed, but the structure is complex, and the thickness is approximate. Inasmuch as exposures in the Pine Creek area are sparse and small, the member cannot be described in detail. However, units of well-sorted light-gray quartzite, which occur mostly in the upper 600 m of the member, have been mapped (Hobbs and others, 1965, pl. 1). The quartzites are nearly matrix-free, but they contain argillite grains that probably were originally rock fragments (fig. 10). The quartzites are clearly lenticular and are generally less than 15 m thick. Lateral dimensions cannot be determined, but one lens extends at least 5 km northward along the valley wall. One quartzite unit is 45 m thick, but it intertongues with siltite-argillite and may be a composite of several lenses.

One quartzite lens is well exposed opposite the mouth of Nabob Creek (Hobbs and others, 1965, pl. 1). The lower 3.7 m of this exposure consists of cross-laminated wavy-bedded (Reineck and Wunderlich, 1968) quartzite in beds 5–10 cm thick separated by argillite layers. The quartzite layers increase in thickness upward. The wavy-bedded quartzite is overlain by 7.6 m of quartzite in indistinct crossbedded sets about 0.3 m thick. This quartzite resembles, in vertical sequence, sandstone bodies ascribed to storm-driven migration of shelf sand ridges by Swift and Rice (1984, p. 48) and Tillman and Martinsen (1984, fig. 9).

Trough crossbedding in another quartzite lens, this one nearly 4 m thick, is illustrated in figure 11. Figure 12 is a graphic section of an exposure near the top of member E on the east bank of Pine Creek just north of the mouth of Amy Gulch (Hobbs and others, 1965, pl. 1). Note the contorted siltite-argillite beds beneath two of the quartzite units. The contortion probably resulted from loading by the migrating sand waves.

The contact between members E and F at the one locality in the Pine Creek area where I have seen it is gradational through about 1.5 m; thinly and irregularly laminated siltite and argillite grade upward into planar-laminated argillite and minor siltite typical of member F.

I have not measured many directional features in member E. The orientations of flute casts at the base of the member east of Kellogg are shown on plate 2 as are the dip directions of cross-lamination at the base

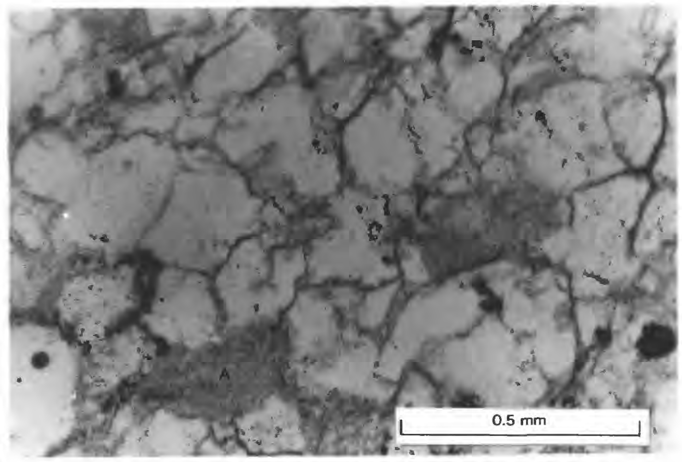


FIGURE 10.—Photomicrograph of quartzite of member E of Prichard Formation. A, argillite grains. From Pine Creek area, Idaho.

FIGURE 11.—Trough crossbedding in quartzite of member E of Prichard Formation in Pine Creek area, Idaho.

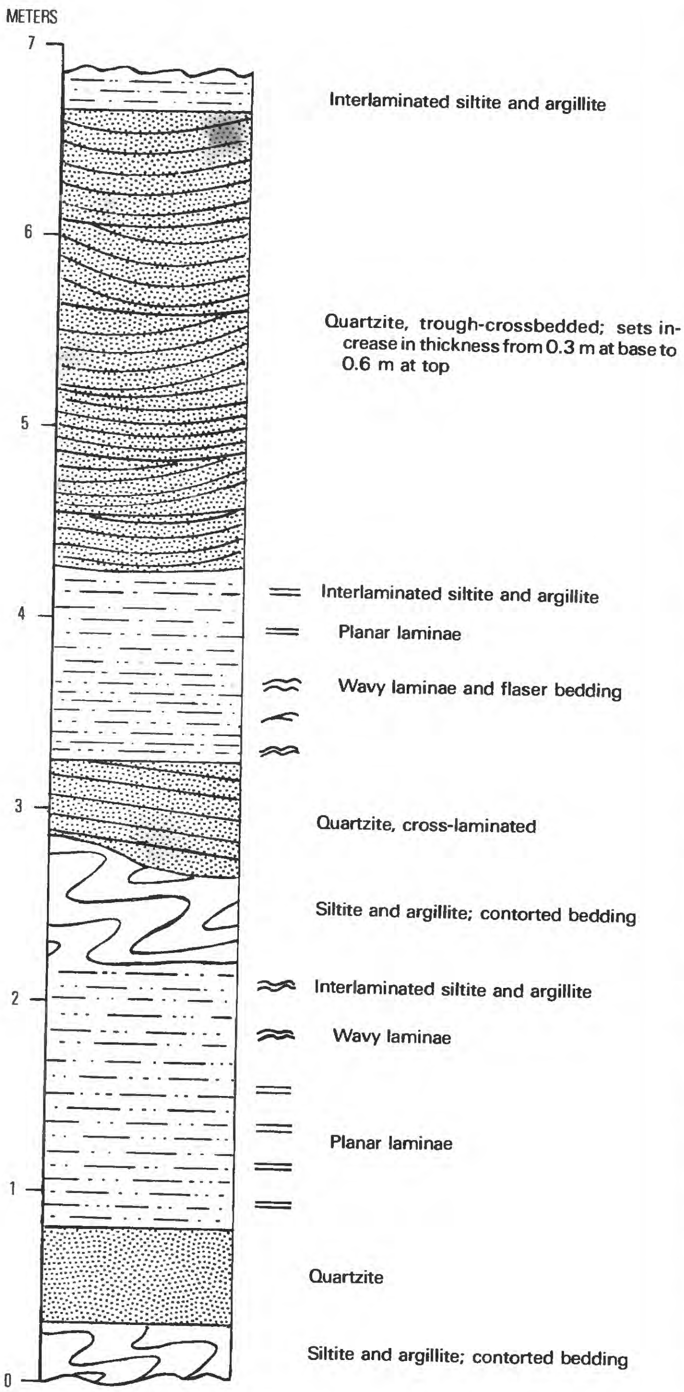


FIGURE 12.—Graphic section of exposure near top of member E of Prichard Formation on east bank of Pine Creek southwest of Kellogg, Idaho.

of one of the quartzite lenses in the Pine Creek area, but the flute casts are too few to give the sense of sediment transport. The lined area on plate 2 is interpreted as a large shelf-edge slump (Cressman, 1985, p. 44–48, fig. 25), and the arrows indicate the direction of slumping inferred from the orientation of fold axes in the slump sheet and the shape of the base of the sheet.

MEMBER F

Member F is composed of planar interlaminated and interbedded light-gray argillite and dark-gray silty argillite to argillic siltite beds that lie between the well-sorted quartzite and irregularly interlaminated siltite and argillite beds of member E below and interbedded quartzite, siltite, and argillite beds of member G above. The distribution and thickness of the member are shown on plate 2. In the area from Kellogg to Coeur d'Alene the distribution of member G is not well known, and in part of the area it is absent, having pinched out; therefore, member F and the lithologically similar member H are not differentiated. An argillite and siltite section that occurs beneath a quartzite-rich section in the outcrop areas east of Chewelah, Wash., is assigned to member F, though the relation of the rocks to the member as identified elsewhere is unknown.

The thicknesses shown on plate 2 for member F in the outcrop area southeast of Plains are somewhat less than those given in Cressman (1985, p. 28), partly because the thickness given in Cressman (1985) includes a sill that intrudes the member and partly because the thicknesses on plate 2 are site specific.

The following description of member F in the exposures east and southeast of Plains from Cressman (1985, p. 28, 29) is equally applicable to the other exposures of the member:

> Typical exposures of member F consist of interlaminated and very thinly interbedded silty argillite and argillite. The two may form graded couplets, or the contacts may be sharp. Most commonly the layers range from 1 to 6 cm thick, and the rock has a banded appearance. The silty argillite is generally dark gray and the argillite medium to light gray, but in parts of the section the silty layer is the lighter of the two. Most laminae and beds are planar, but a few silty beds pinch and swell and channel the underlying argillite. Many silty argillite layers are internally laminated; the laminae consist of silty argillite-argillite couplets, concentrations of carbonaceous matter within otherwise uniform argillite, or pyrite. Pyrite as discontinuous laminae generally less than 1 mm thick and as grains strung along bedding is characteristic of member F. Though the discontinuous pyrite laminae are not present in all beds, I have seen them in all parts of the member. Inasmuch as similar pyrite laminae are also present in the lower part of member H, the presence of the laminae, though very suggestive of member F, is not diagnostic. A few argillite beds are as thick as 0.5 m and contain few or no silty laminae. These weather to light-gray massive-appearing outcrops that are particularly characteristic of the lower part of the member.

Much of the material described as silty argillite above contains at least 40 percent quartz silt, and some might more appropriately be termed argillic siltite; the distinction is difficult to make in thin section and nearly impossible to make in hand specimens.

The iron sulfide, which elsewhere may be pyrrhotite rather than pyrite, weathers to impart to the rock the dark-reddish-brown hues that are a distinguishing feature of much of the Prichard Formation (Ransome and Calkins, 1908, p. 32).

The argillites of member F probably represent fine-grained turbidites and suspended-load deposits.

Member F in the exposures east and southeast of Plains contains several argillite-pebble conglomerates of uncertain but local extent (Cressman, 1985, p. 25–30). One of these conglomerates is well exposed in a roadcut for Montana Highway 135 just south of Siegel Creek 17 km south-southeast of Plains. This conglomerate, which is nearly 24 m thick, consists of three units of matrix-supported argillite-pebble conglomerate and two intercalated units of very poorly sorted argillic siltite. Pebbles in the conglomerate are oriented subparallel to bedding, and drag features are present in the underlying bed. These conglomerates nearly certainly resulted from thixotropy and flow of an original sediment of interlaminated silt and clay. The poorly sorted argillic siltites resulted from nearly complete disaggregation and mixing, whereas the conglomerate layers resulted from less severe thixotropy in which the silty layers were completely disaggregated whereas the clayey layers were broken into fragments. The disruptions of the sediment may have resulted either from surface flow or from slip along a zone within the sediment. The local extent of the conglomerates suggests surface flow, and the conglomerates probably originated as submarine debris flows.

I have not seen silt-pebble conglomerate in the other outcrop areas of member F, but light-gray argillite beds similar to those described in the lower part of the member in the quotation above are present wherever the lower part is exposed. For example, east of Kellogg, where the contact between members E and F is well exposed, the basal part of member F consists in ascending order of 20 cm of interbedded planar light-gray argillite and dark-gray silty argillite, 45 cm of structureless light-colored silty argillite, 20 cm of interbedded planar argillite and silty argillite, 62 cm of structureless silty argillite, and many meters of interbedded planar argillite, silty argillite, and siltite. Another example, from the Pine Creek area, is shown in figure 13. A few of the light-colored beds contain faint convolute bedding that resulted from plastic flow, but the majority of beds are structureless and were apparently either deposited rapidly from suspension or from fluidal flow that disrupted all internal structure.

Black cherty-appearing tourmalinized siltite occurs in member F in several localities. About 3.2 km southeast of Siegel Mountain, southeast of Plains, Mont., a tourmalinized bed, possibly as thick as 1 m, crops out about 315 m above the base of the member and about 50 m above a diorite sill. The tourmalinized bed extends several kilometers east-southeast along strike (R.E. Van Loenen, written commun., 1981). Similar cherty-appearing tourmalinized rock occurs as float near the whitewashed letter H on the hillside northeast of Hot Springs, Mont., where the tourmalinized rock is adjacent to a diorite sill. Tourmalinized rock has also been reported by J.P. Thorson (written commun., 1985) from several localities in member H about 10 km northeast of Murray, Idaho. In this area, member H contains no sills.

FIGURE 13.—Massive argillite near base of member F of Prichard Formation in Pine Creek area, Idaho. Note massive beds separated by zones of indistinct planar bedding.

In the one thin section of tourmalinized rock from member F that I have examined, the tourmaline consists of a felted mass of acicular crystals that are less than 1 μm in diameter and several micrometers in length. The tourmaline has preferentially replaced the argillite in interlaminated and very thinly interbedded argillite and siltite and is concentrated along foresets of cross-laminae.

Tourmalinized rock similar to that described above occurs in the Aldridge Formation in the footwall of the Sullivan orebody of British Columbia (Ethier and Campbell, 1977, p. 235). The tourmaline there is rich in magnesium and intermediate in composition between dravite and schorl. Ethier and Campbell (1977, p. 2359, 2360) infer the following origin: (1) fluids containing boron, either derived from igneous intrusions at depth or scavenged from the sedimentary pile, discharged through fractures created by sea-floor collapse; (2) these fluids concentrated within local basins; (3) boron-rich and boron-poor layers were formed syngenetically in the sediments; and (4) subsequent metamorphism modified the composition of the resultant tourmaline.

The tourmalinized rock in member F has excited considerable interest because tourmaline is a common gangue mineral in many strata-bound massive sulfide deposits (Slack, 1982, tables 1, 2; Taylor and Slack, 1984, p. 1704) and because tourmalinization is widespread and abundant in footwall rocks of the Sullivan orebody in the Aldridge Formation of British Columbia (Freeze, 1966, p. 272; Ethier and Campbell, 1977).

DOLOMITIC SILTITE MEMBER

In the Gibbs No. 1 borehole the interval between member E below and member G above consists of dolomitic siltite instead of the planar interbedded argillite and siltite that occupies that stratigraphic position to the south and southwest. Dolomite makes up as much as 50 percent of some beds, and both biotite and microcline porphyroblasts are common. The unit, informally termed the dolomitic siltite member, is 920 m thick.

The only place where the dolomitic siltite member is exposed is on the east side of Upper Whitefish Lake 33 km north-northwest of Whitefish, Mont. (pl. 1, sec. 15). Here, very thinly and planar interbedded light-gray argillite and dark-gray laminated silty argillite of member H grades downward into planar interlaminated pinkish-gray to light-brownish-gray dolomitic siltite in layers 1–10 mm thick and medium-gray argillite in layers 0.5–3 mm thick. Most laminae are in sharp contact, but a few form graded couplets. The interlaminated dolomitic siltite and argillite grade downward into massive medium-gray dolomitic argillite that contains some biotite porphyroblasts and some equant grains of pyrrhotite. The rock weathers brownish gray. Microcline porphyroblasts that are common in the unit in the Gibbs No. 1 borehole are not obviously present. The total thickness of the exposure probably does not exceed 40 m. The correlation of this exposure with the dolomitic siltite in the Gibbs No. 1 borehole as shown on plate 4 is reasonable but by no means certain.

MEMBER G

Member G, first mapped and described in the outcrops east and southeast of Plains, Mont. (Cressman, 1981; 1985, p. 33, 34), consists of 500 m of interbedded quartzite, siltite, and argillite between banded light-gray and dark-gray argillite and siltite of member F below and of member H above. The unit has since been identified as far north as 40 km south of Eureka, Mont., and as far west as 15 km east of Coeur d'Alene, Idaho. The thickness ranges from more than 1,130 m near Bear Creek north-northwest of Thompson Falls to a feather edge between Kellogg and the Pine Creek area. Quartzite interbedded with siltite and argillite occurs in about the same stratigraphic position as member G in exposures of dynamically metamorphosed Prichard Formation that occur on thrust plates near the southern border of the report area, but I do not know whether these rocks were once continuous with member G or whether they are part of a separate quartzite-bearing unit. The distribution and thickness of member G are shown on plate 2.

Member G is a tongue of the quartzite facies that extends well into the argillite facies; therefore, the lithology of member G will be discussed together with that of the quartzite member of the quartzite facies rather than being presented separately here.

MEMBER H

A section of planar, very thinly interbedded argillite, silty argillite, and siltite that overlies the quartzite of member G and underlies irregularly interlaminated siltite and argillite of the transition member was assigned to member H of the Prichard Formation in the area from Plains to Perma, Mont. (Cressman, 1981; 1985, p. 34–37). The member has been recognized northward to the west side of Glacier National Park and westward to near Coeur d'Alene, Idaho (pl. 2). From Kellogg west to Coeur d'Alene the distribution of member H is not shown separately on plate 2 because member G has been mapped only locally. Also, the Pine Creek area is south of the pinch-out edge of member G, and in the absence of that member, members F and H cannot be differentiated. The argillite member and the upper member of the quartzite facies are tongues of member H.

The thickness of member H in the Plains-Perma area is uncertain because of faulting and inadequate exposures, but map measurements near Deemer Peak (Cressman, 1981), where the member is fairly well exposed and there is no evidence of internal faulting, give a thickness of 1,450 m. This is about twice as thick as in other occurrences to the west where the member is 580–730 m thick, but map relations do not permit a thickness east of Plains of less than that at Deemer Peak.

In Glacier National Park the base of member H is not exposed. According to Earhart and others (1983) the thickest section of the member, which they called the Prichard Formation, is about 850 m thick. Measurement from their map indicates a greater thickness.

The basal part of member H is indistinguishable in lithic character from member F and consists of interlaminated and very thinly interbedded medium- to light-gray argillite and dark-gray silty argillite. The dark-gray silty argillite commonly contains irregular, discontinuous laminae of carbonaceous matter and iron sulfide. The beds are planar and most contacts are sharp, though there are some graded couplets (fig. 14). The alternating light-gray and dark-gray planar layers give the rock a banded appearance (fig. 15). Higher in the section the argillite layers thicken and the silty argillite layers thin, and argillite makes up two-thirds to three-fourths of the member. The general appearance changes upsection from banded to lined; that is, on an exposed face, the silty layers appear as straight lines ruled across the argillite surface. In the member as a whole, silty argillite and siltite make up only one-fourth to one-third of the section. A few of the silty layers are lenticular and cross-laminated, and some are argillaceous siltite rather than silty argillite. These cross-laminated units tend to be lighter color than the argillites. The entire member weathers to the same rusty color as member F.

The exposures of member H west and southwest of Patricks Knob (south of Plains, Mont.) contain a unit 90- to 120-m-thick of interbedded siltite, quartzite, and argillite, the base of which is about 610 m above the base of the member. Cross-lamination, small-scale scour-and-fill structure, and ripple marks are present in the siltite and quartzite. The ripple marks are asymmetric, have gently curved crests, and commonly end in "tuning forks." The amplitude is 5 mm and the wavelength is 1–3 cm. Considering compaction from an original porosity of 56 percent (Watts, 1981, fig. 8) and a present porosity of near zero, the original amplitude would have been about 11 mm. The planar view of the ripple marks suggests that they were formed by wave current or combined flow as described by Allen (1982, p. 429–431). This unit is present both north and south of the St. Marys fault in this vicinity (Cressman, 1981), but I have not found it elsewhere, even in the other exposures in the Plains area.

Southeast of Thompson Pass, member H contains a unit that consists of beds of structureless greenish-gray silty argillite. This unit may be widespread in the Coeur d'Alene district.

Slump folds are present in member H but are not abundant except in Glacier National Park where slump-folded units, mostly about 0.3 m thick, are common along McDowell Creek. The exposure in the Park also has, near its base, a bed that contains elongate argillite clasts, a few rounded dolomite grains about 0.3 mm in diameter, rounded quartz grains as much as 0.6 mm in diameter, and a trace of microcline, all set in a matrix of silty argillite. The bed is probably a debris flow. Also in the Park, member H, near its top, contains lenses and nodules of manganiferous carbonate (J.W. Whipple, oral commun., 1984).

Like member F, member H probably contains both fine-grained turbidite beds and beds deposited from

FIGURE 14.—Interbedded argillite and silty argillite of member H of Prichard Formation exposed in McDonald Creek, Glacier National Park. Light-gray argillite interbedded with dark-gray planar-laminated silty argillite. Note contorted bedding immediately below knife.

FIGURE 15. Photomicrograph of graded siltite-argillite couplets of member H of Prichard Formation. Siltite, light gray; argillite, dark gray. Black irregular laminae near base consist of carbonaceous matter. Sample from McDonald Creek in Glacier National Park.

suspension; turbidite beds are probably the most abundant.

The contact of member H with the overlying transition member of the Prichard Formation can generally be located to within a meter in adequate exposures. In most localities argillite that contains continuous planar silty laminae typical of member H is overlain by silty argillite in structureless beds as much as 1 m thick that alternate with thin sets of alternating planar to wavy and nonparallel very thin beds of siltite and argillite. The siltite may be cross-laminated and ripple marked. Locally, the basal bed of the transition member consists of slump folds that are outlined by faint laminae in an otherwise massive bed.

A description of the transition member follows the descriptions of units of the quartzite facies.

QUARTZITE FACIES

The subdivisions of the quartzite facies are based on mapping in the Yaak River area in the northwest corner of Montana (Cressman and Harrison, 1986). The Prichard Formation there is represented by section 11 on plate 4.

QUARTZITE MEMBER

The section of interbedded quartzite, siltite, and argillite that makes up most of the quartzite facies is informally termed the quartzite member. The term is appropriate in that gray quartzite beds form about 65 percent of the unit. Member G is a tongue of the quartzite member that extends into the argillite facies and will not be described separately. Locally, other units are intercalated in the member, and sills are present throughout most of its area.

The distribution of the quartzite member and its thickness, exclusive of the intercalated members and the sills, are shown on plate 2. The member thickens from a feather edge somewhere between the Gibbs No. 1 borehole and Upper Whitefish Lake, Mont., and between Kellogg and the Pine Creek area, Idaho, to a maximum thickness of at least 6,000 m north of Newport, Wash. At Newport the base of the member is not exposed. From there the member thins westward to about 1,300 m east of Chewelah, Wash., where both the upper and lower contacts are exposed.

The quartzite member consists of quartzite-dominant intervals that alternate with siltite-argillite intervals. The quartzite-dominant intervals average 10–15 m thick but range from 2 to 45 m. The siltite-argillite intervals average about 8 m thick but range from 2 to 20 m. This character is illustrated by the stratigraphic sections on plate 3 and by figure 16. These intervals are thicker in the exposures west of Troy than they are in the Yaak River area north of Troy. Furthermore, in the exposures west of Troy the intervals are thicker to the south than to the north. Note, for example, that on plate 3 section C contains a siltite-argillite interval 35 m thick and that section A contains a quartzite interval 45 m thick.

Individual quartzite beds within the quartzite intervals are turbidites as first recognized by Bishop and others (1970). The quartzite beds have sharp bases, some of which are erosive, and nearly all grade to argillite, siltite, or very fine grained quartzite in their top few centimeters. Dewatering vents present locally (fig. 17) indicate that deposition was episodic; that is, a freshly deposited bed had time to dewater and compact before deposition of the overlying sediment. Basal

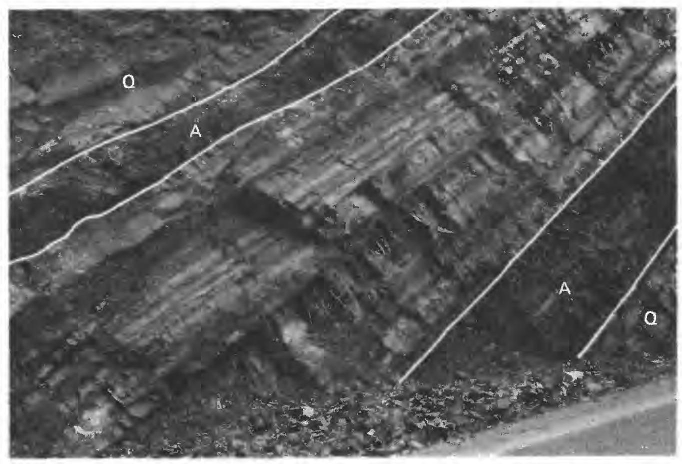


FIGURE 16.—Part of quartzite member of Prichard Formation exposed in cut along Katka Face road east of Bonners Ferry, Idaho. Note alternate intervals of predominant quartzite (Q) and predominant argillite (A).

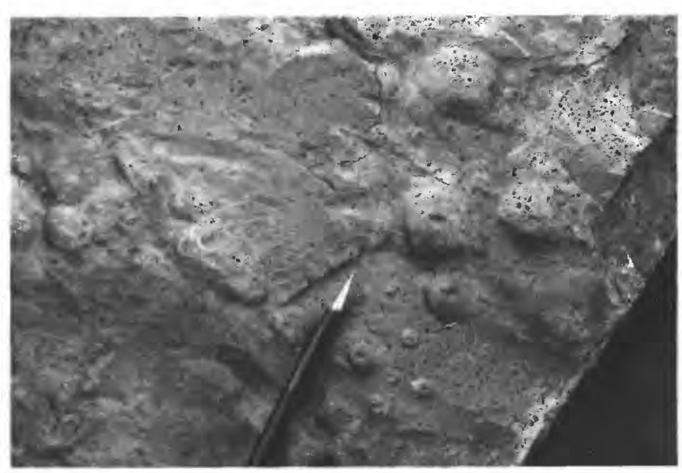


FIGURE 17.—Dewatering vents on top surface of quartzite turbidite in quartzite member of Prichard Formation in cut on Katka Face road east of Bonners Ferry, Idaho. Black specks are asphalt from road surface.

contacts are mostly planar, but a few beds channel several centimeters into the underlying bed, and some exhibit load casts, flute casts, groove casts, and ridge-and-furrow casts at their base (fig. 18). Tool marks are absent because metazoans, whose hard parts are commonly the tools, had not evolved.

Structures in sandstone turbidites are commonly described in terms of the Bouma sequence. Bouma (1962) described an ideal turbidite bed as consisting of five structural divisions designated A–E from base to top. A is massive or graded, B is parallel laminated, C is cross-laminated or convoluted, D is parallel laminated, and E is pelagic shale. In the quartzite member most of the beds consist of massive quartzite that grades upward to siltite and argillite. These were termed Bouma AE turbidites by Edmunds (1973). The Bouma D division is present at or near the top of some beds as parallel-laminated siltite and argillite. The Bouma C division is present in a few beds, but the B division is rarely recognizable. Rip-up clasts are present near the base of a few beds. I cannot document any regional or consistent stratigraphic variation in the Bouma sequences; if such variations exist, they are too subtle to have been noted in this reconnaissance study.

The median thickness of the quartzite-turbidite beds in all the sections on plate 3 is 0.44 m (fig. 19), though the mode is between 0.2 and 0.4 m. The maximum thickness is 7 m, but 95 percent of the beds are less than 2 m thick and 80 percent less than 1 m thick. This general uniformity of bedding thickness is illustrated by the exposure in figure 20. Thick beds of amalgamated quartzite are present, but individual turbidites can generally be distinguished on close inspection.

In the measured sections on plate 3 the median thickness of the quartzite-turbidite beds is as follows:

Section	Median thickness (meters)
A	0.9
B	0.65
C	0.35
D	0.4
E	0.23
F	0.4
G	0.55
H	0.45
I	0.55

Sections A and B, the most southwesterly of those measured, are thicker bedded than the other sections; otherwise no trend is apparent, probably because the sections are not correlative.

Within the quartzite intervals many of the beds are arranged in thickening-upward sequences that are mostly 1–2 m thick, though some are as much as 7 m thick

FIGURE 18.—Flute and groove casts at base of quartzite turbidite in quartzite member of Prichard Formation exposed in cut on Pete Creek road north of Yaak River, Mont. See Cressman and Harrison (1986) for location.

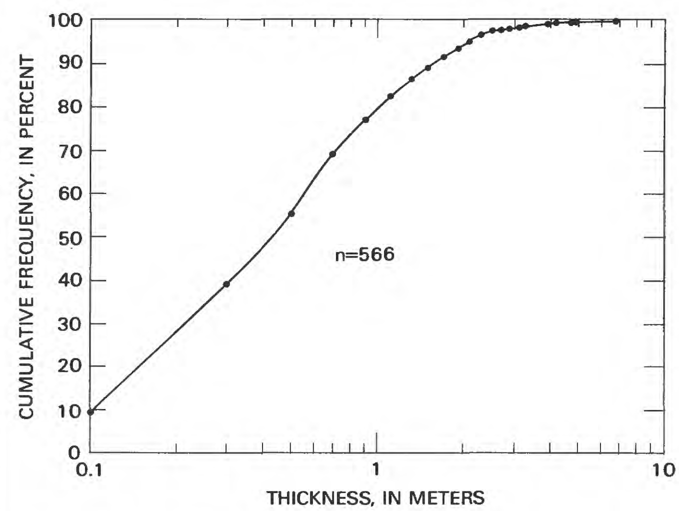


FIGURE 19.—Thickness frequency of quartzite-turbidite beds in quartzite member of Prichard Formation.

and one (pl. 3, sec. A) is 11 m thick. Thinning-upward sequences are also present but are less common; they are mostly 1–3 m thick. However, many beds do not seem to be part of any regular vertical trend. Plate 3 illustrates these characteristics.

Sparse, nearly spherical to oblately spheroidal concretions 0.05–0.3 m in diameter are present in some quartzite beds. The concretions are defined by their dark color, which is nearly black in the center, and by their tendency to weather out more rapidly than the adjacent rock. The concretions themselves are composed of quartzite, but I do not know what differences in composition differentiate them from the surrounding quartzite. One specimen from the nearly black interior of a concretion contained abundant minute tourmaline

FIGURE 20.—Exposure of quartzite turbidite beds of quartzite member of Prichard Formation exposed in Moyie River 18 km northeast of Bonners Ferry, Idaho.

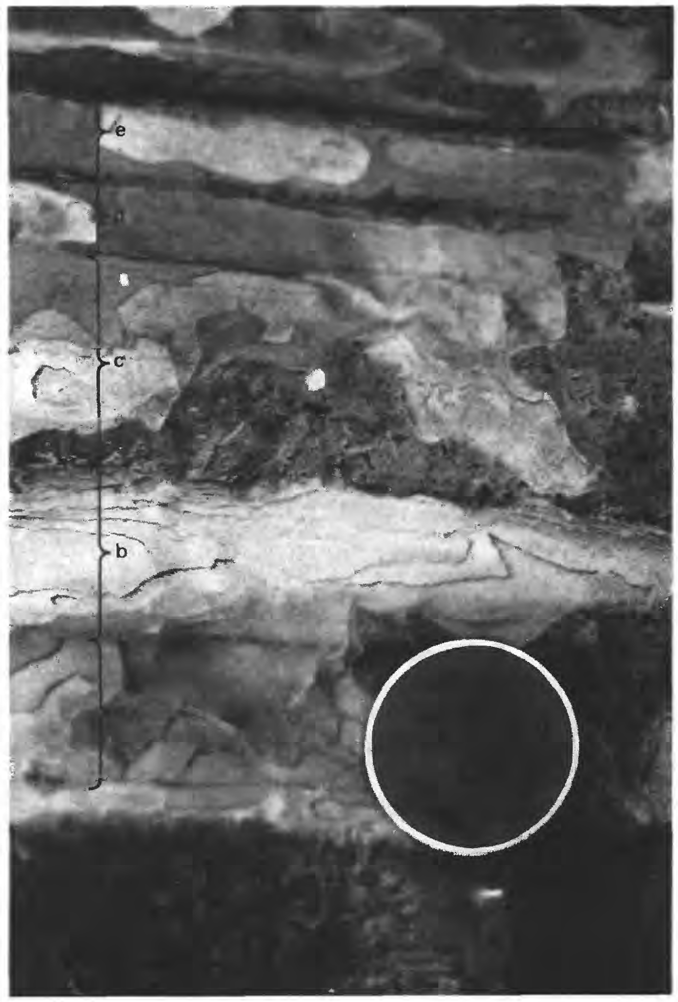


FIGURE 21.—Graded siltite-argillite couplets in quartzite member of Prichard Formation; a through e are individual couplets. Note parallel-planar lamination in couplet c. Siltite forms basal one-fourth to one-third of each couplet. Lens cap is 5.2 cm in diameter. Exposed along North Fork of Callahan Creek west of Troy, Mont.

crystals, but I do not know if this is a unique occurrence or a feature common to most concretions. The concretions are present throughout the section wherever quartzite turbidites are present.

The siltite and argillite intervals in the quartzite member consist of graded siltite-argillite couplets, very thinly interbedded and interlaminated siltite and argillite, and laminated light-gray and dark-gray argillite. The graded siltite-argillite couplets range from about 2 mm to as much as 17 cm thick, though most are 0.5–2 cm thick. Figure 21 illustrates some of the thicker couplets. The couplets may consist mostly of siltite, mostly of argillite, or of about equal proportions of each. The thicker couplets, such as those in figure 21, are, in general, mostly of argillite. The basal silty part is most commonly either structureless or planar laminated, but cross-laminae are present locally. These laminae commonly consist of alternating, more and less silty, layers about a millimeter thick, though some consist of irregular, discontinuous concentrations of carbonaceous matter. The upper argillaceous part may also be marked by planar-parallel laminae. A few couplets contain rip-up clasts of argillite in their basal part. Irregular layers and lenses of siltite, some of which are cross-laminated, may be interleaved in sequences of graded couplets.

The graded siltite-argillite couplets, like those of members A and B, were probably deposited from low-density turbidity currents. The graded couplets closely resemble the fine-grained turbidites described by Stow and Shanmugan (1980), and according to their terminology the most common structural sequence in the fine-grained turbidites of the quartzite member consists of divisions T_3–T_7; sequences of T_1–T_7 and T_0–T_7 are much less common. The graded couplets are thinner than most fine-grained turbidites described in the literature, but assuming an original porosity of 56 percent (Watts, 1981, fig. 8) and a present porosity of near zero, the original thicknesses have been reduced at least 50 percent by compaction.

The interlaminated and very thinly interbedded siltite and argillite that make up part of the siltite-argillite intervals in the quartzite member differ from the siltite-argillite couplets in that all contacts are sharp. The siltite is argillaceous and dark gray; the color results from abundant biotite and, locally, carbonaceous laminae. The layers are planar and range from less than a millimeter to several centimeters thick.

The least common rock type in the siltite-argillite intervals of the quartzite member is interlaminated light-gray and dark-gray silty argillite. The laminae are planar and exhibit no obvious current effects. The dark laminae contain discontinuous carbonaceous laminae and, locally, discontinuous sulfide laminae; otherwise, the light and dark laminae do not differ in composition. Sets of the laminated argillite are rarely as much as 1 m thick. They commonly occur as thin sets within a sequence of siltite-argillite couplets or, less commonly, as thin sets between quartzite beds. An outcrop of the laminated argillite is illustrated by figure 22.

Geologists of Cominco Ltd., which operates the Sullivan mine in British Columbia, have used 14 of the thicker laminated argillites as stratigraphic markers and have traced some of them nearly 300 km by matching the patterns of lamination (Edmunds, 1977b, p. 22). Huebschman (1973) traced and described one of these marker units in considerable detail. In the Yaak River area I locally mapped six intervals of the laminated argillite (Cressman and Harrison, 1986). The stratigraphic position of these units is shown on column 11, plate 4, as key beds a through f. These markers, which were also identified locally in the exposures west of Troy, are also shown on section 10, plate 4.

Both Huebschman (1973, p. 697) and Edmunds (1977b, p. 22) attributed the laminated argillite to deposition in pelagic (or hemipelagic) environments. The wide extent and uniformity of these beds makes any other explanation unlikely. Huebschman (1973, p. 697) considered the organic matter to have been derived from blooms of plankton and the silt to have been derived from dust storms. I am not as impressed as Huebschman by the amount of silt in the marker zones, and settling from suspension of fines contributed by rivers would seem an adequate explanation for the detrital component.

The siltite-argillite intervals of the quartzite member locally exhibit slump folds, but they are uncommon. The thickest slump-folded unit that I have observed is shown in section A, plate 3, and is 13 m thick. The overlying 10 m of quartzite is structureless and may also have slumped or flowed, so the entire slump complex may be 23 m thick. The directions of slumping, which were determined in the same manner as for the slump folds of member B, are shown on plate 2. Note that many folds are at a high angle to the direction of transport determined from sole marks on the quartzite turbidites.

Several submarine channels as much as 2 m deep and filled with quartzite turbidites are present in the quartzite member in the dynamically metamorphosed exposures west to north-northwest of Bonners Ferry, Idaho. Elsewhere, evidence for submarine channels and channel fill is very sparse. One exposure on the Katka Face road 11 km east of Bonners Ferry displays one sequence of quartzites resting in sedimentary contact on another with an angular discordance of 20°, and the upper beds are probably channel fill. Large channels—those hundreds of meters wide and tens of meters deep—probably could not be detected in the available exposures.

FIGURE 22.—Laminated silty argillite in quartzite member of Prichard Formation. This is key bed "d" mapped by Cressman (Cressman and Harrison, 1986) and is exposed in bank of Spread Creek directly north of Yaak River road. See Cressman and Harrison (1986) for location.

About 11.5 km north of Clark Fork, Idaho, on a National Forest road to Porcupine Lake, a quartzite more than 10 m thick that contains argillite fragments laterally abuts argillite and siltite at a high angle. The contact between the quartzite and siltite-argillite may have been a fault penecontemporaneous with sedimentation or it may have been the contact of levee with channel fill on the outside of a channel meander.

An unusual breccia is exposed in the quartzite member at an altitude of 5,200 ft (1,585 m) on the road up Smith Mountain west of Troy (Mt. Pend Oreille 15-minute quadrangle). Smith Mountain is the location of section B on plate 3. The bedding there strikes north-south about normal to the trend of the mountain side and dips about 25°–35° to the east. A breccia body is exposed at road level; it is concealed to the west but is in vertical contact to the east with interbedded quartzite and argillite. The breccia extends upward to about 45 m above the road where it is capped by bedded siltite and argillite; the extent of the breccia below the road is not known. Relations up the hill, though not entirely clear, suggest that the breccia body is nearly vertical, about 20 m wide, and pipelike or dikelike in shape. The eastern contact cuts across bedding in the surrounding rock. The breccia itself consists of elongate clasts of argillite and minor quartzite as much as 2.5 cm in length that float in a matrix of very poorly sorted argillaceous quartzite. Elongate clasts are oriented and plunge 67° S. 50° E. At two localities where the nearly vertical contact with bedded rock is exposed, the adjacent beds dip about 60°—about 30° greater than the dip at some distance—and are interleaved with breccia at a scale of several centimeters to several tens of centimeters. Two breccia beds several meters thick, one about 10 m above the road and the other at the top of the vertical breccia mass, extend eastward from the breccia dike or pipe parallel to bedding for a distance of at least 200 m. The breccia in these beds is similar to that in the vertical mass, but the clasts are not obviously oriented.

The breccia is probably a dewatering feature. This is most clearly indicated by the increased dip adjacent to the pipe or dike which indicates upward movement of the breccia. Furthermore, the breccia interleaved with unbroken bedded rock at the margin of the vertical breccia mass suggests that the breccia was fed from the adjacent country rock. I have seen no evidence to indicate whether the two breccias concordant with bedding were spread on the surface from a vent or whether they were injected as sill-like bodies.

The contact between the quartzite member and argillite of the overlying upper member, or in the case of member G, argillite of member F, is gradational in most localities as quartzite sets become thinner and more widely spaced upward. In most localities I have placed the contact at the top of the uppermost quartzite, and as a result the uppermost part of the quartzite member may consist mostly of argillite. The zone of gradation is generally thin considering the scale of the graphic sections, but in the section north of Newport (pl. 4, sec. 12) the uppermost 500 m shown as the quartzite member consists mostly of argillite and siltite. Similarly, in the Bear Creek section (pl. 4, sec. 7) the upper half of the interval identified as member G is mostly siltite and argillite. In more detailed work the members might be shown to intertongue.

MASSIVE MEMBER

The massive member crops out both north and south of the Yaak River in an area of 50 km^2 that is centered about 10 km north of Sylvanite, Mont. (fig. 2, pl. 2) where the member is intercalated in the lower part of the quartzite member (pl. 4, sec. 11) (Cressman and Harrison, 1986). The base of the massive member is 550 m above the base of the exposed section and about 5,100 m below the top of the Prichard Formation. The full extent of the member is not known, but it is not present in the exposures between Troy and Bonners Ferry where that part of the section is again exposed. The thickness of the member is also uncertain, but it probably ranges from 150 to 250 m.

The massive member consists of medium-dark- to medium-light-gray very poorly sorted argillic quartzite that weathers a characteristic light brown to brownish gray. The rock is massive without any indication of bedding and crops out as thick rounded ledges.

The member contains scattered matrix-supported plate-shaped bodies of silty argillite that are as large as 70 cm in diameter but are more commonly 20–30 cm across and several centimeters thick. They are mostly exposed in section as planar to gently curved lath-shaped bodies (fig. 23). The plates are mostly randomly oriented, but locally they may be preferentially oriented vertical, inclined, or horizontal to bedding as inferred from the subjacent and superjacent section. Some of the plates, such as those in figure 23, have light-colored rinds that surround a dark, biotite-rich interior. The argillite plates are not uniformly distributed throughout the member, but I was unable to detect any regular stratigraphic variation in either their distribution or their size. Rounded quartzite fragments are also present but not in the abundance of the argillite plates.

The massive member everywhere rests directly on a sill that is about 60 m thick, and a few argillite plates are present in the uppermost part of this sill. In some localities the member is directly overlain by a sill; in others, a thin sill occurs within the upper part of the member.

The presence of the sill at the base of the member suggests a causal relationship between the sill and the member. I suspect, but cannot prove, that the unique features of the massive member resulted from sill intrusion, the conversion of pore water in the sediments to steam, a resultant increase of pore pressure, and consequent fluidization of the sediment. Sea-floor slumping, an alternate explanation, could not explain the presence of argillite plates in the top of the sill;

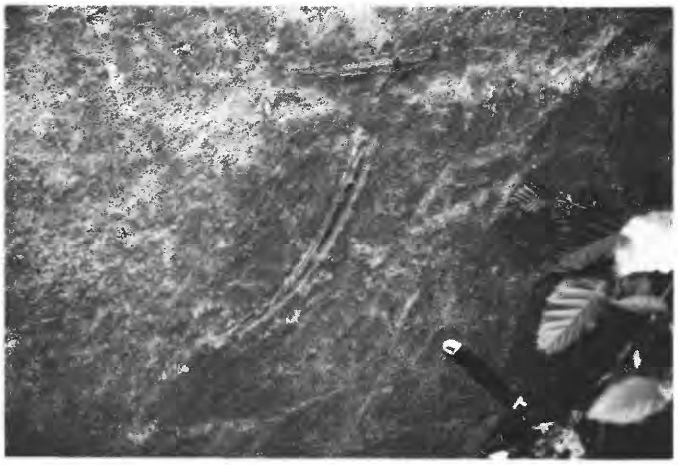


FIGURE 23.—Exposure of massive member of Prichard Formation showing lath-shaped sections of platelike argillite bodies and lack of bedding. Exposure is orthogonal to regional bedding.

furthermore, considerable rapid down-slope movement and turbulent flow would have been required to destroy all signs of bedding and to thoroughly mix the constituents of the original interbedded clay silt and sand, but the argillite plates are only mildly deformed.

ARGILLITE MEMBER

I applied this name (Cressman and Harrison, 1986) in the Yaak River area north of Troy, Mont., to a unit about 240 m thick that is composed mostly of argillite and occurs 1,660 m below the top of the Prichard Formation (section 11, pl. 4). The unit is both underlain and overlain by the quartzite member of the Prichard. The thickness and distribution of the member are shown on plate 2. More detailed work may show that the member extends farther west than shown on the plate.

The argillite member is the same unit described by Huebschman (1973, p. 689–693) as unit C. Huebschman described a laminated siltite unit (identified as argillite in this report) 4 m thick within the member that is one of the marker beds identified by Cominco geologists. Huebschman traced both the marker and the member into Canada for a distance of 65 km north of the international border.

The argillite member consists mostly of argillite and silty argillite that is laminated and medium light gray and medium dark gray. The fresh rock is very resistant to impact and contains abundant pyrrhotite, the weathering of which imparts a rusty color to the rock. The pyrrhotite is so abundant that an adit to explore for sulfide ore was driven in the member on the west side of the Yaak River 8 km south (downstream) of Sylvanite. The argillite weathers to even slabs 0.3–0.5 m thick.

The lower part of the argillite member contains some argillite and laminated silty argillite interbedded at a scale of several centimeters. The lower part also commonly contains several beds about 0.5 m thick that either lack internal structure or exhibit faint laminae that indistinctly outline slump folds.

The upper two-thirds of the member consists of argillite that is laminated medium light gray and medium dark gray. The dark laminae consist of alternating layers of carbonaceous matter and pyrrhotite lenses. Some dark laminae contain more silt and biotite than the light-colored laminae; others do not. The light- and dark-colored laminae are commonly grouped into bands several centimeters across. Nearly all the laminae are planar, and nearly all contacts between laminae are sharp. Some small-scale scour-and-fill structures and some cross-lamination are present. The member bears some resemblance to much of member F and the lower part of member H of the argillaceous facies.

The marker unit described by Huebschman (1973) is probably a hemipelagic deposit, and some of the other beds in the member were probably deposited by low-density turbidity currents.

The base of the argillite member is placed at the top of the underlying quartzite sequence, and the top is placed at the base of the overlying quartzite sequence. Both contacts are sharp.

UPPER MEMBER

The upper member is a unit of planar-laminated and planar-bedded argillite that lies between the quartzite member below and irregularly interbedded siltite and argillite of the transition member above. The name "upper member" has been widely applied to this unit in the Belt basin, though at times and places the name has included both the member described here and the overlying transition member. On the geologic map of the Wallace 1°×2° quadrangle (Harrison and others, 1986) the name upper member was used at some places for member H and at other places for a unit consisting of member H and the transition member. The distribution and thickness of the upper member as described in this report are shown on plate 2.

The upper member is of medium-light-gray argillite that contains thin, planar layers of light-gray siltite and dark-gray silty argillite. In some exposures perpendicular to bedding, the argillite gives the appearance of having been ruled by light and dark lines (fig. 24), and this has given rise to the term "lined rock" commonly applied to this unit in both the United States and Canada. The medium-light-gray argillite commonly occurs in laminae and beds that range from 1 mm to 5 cm thick but average 1–2 cm thick; the light-gray

FIGURE 24.—Exposure of upper member of Prichard Formation showing medium-light-gray argillite and planar laminae of light-gray siltite and dark-gray argillite. This is "lined rock" of field usage.

FIGURE 25.—Planar and lenticular siltite laminae in argillite of upper member of Prichard Formation. Note faint cross-lamination in some lenticular siltite layers.

siltite and the dark-gray silty argillite are mostly in laminae less than 1 cm thick. Locally, particularly in the lower half of the member, the siltite laminae are absent, the silty argillite and argillite layers are of about equal thickness, and the rock has the banded appearance typical of parts of members F and H of the argillaceous facies of the Prichard Formation. The dark-gray silty argillite contains discontinuous laminae of iron sulfide, mostly pyrrhotite, and carbonaceous matter. The iron sulfide is sufficiently abundant in the upper member between Libby and Kalispell and in the Thompson Falls area that the sulfides form a conductive zone which can be detected by audio-frequency magnetotelluric methods (Wynn and others, 1977; Long and Hoover, 1984; Van Blaricom, 1984). Although planar lamination and bedding is a characteristic and dominant feature of the upper member, irregular and lenticular layers of siltite are common locally (fig. 25), particularly in the upper part.

The argillites and siltites of the upper member were probably deposited partly from low-density turbidity currents and partly from suspension on the basis of the same evidence discussed for the other members of the Prichard Formation that consist mostly of argillite and siltite.

White quartzite lenses that contain garnet, hornblende, and local calcite are a conspicuous, though minor, component of the upper part of the upper member in the exposures east of Libby and west of Kalispell. These lenses occur as far west as the outcrop belt east of Sylvanite. As discussed previously, these beds are probably metamorphosed dolomitic siltstone. Carbonate lenses are present in the same part of the section in the less metamorphosed rocks of Glacier National Park.

At locality A (pl. 2) 35 km S. 80° E. of Libby a quartzite unit 12 m thick and 97 m below the top of the upper member is exposed (the upper member there is about 560 m thick). The base of the quartzite is not exposed. The quartzite is very light gray and very fine grained and is better sorted than the turbidite quartzites of the quartzite member. Beds range in thickness from 0.5 to 3 m but are mostly 0.6–0.9 m thick and are arranged in several indistinct thinning-upward sequences. The beds, most of which grade to siltite in their upward part, are mostly even, but some load casts and some small-scale flute casts are present at their base. The flute casts, though few, indicate a northerly transport direction. This unit does not crop out elsewhere, even though the same stratigraphic horizon is widely exposed. Both its local occurrence, and the thinning-upward sequences indicate that the quartzite fills a channel through which submarine currents transported sand northward.

Slump folds are common in the basal part of the upper member, but the lamination is commonly so indistinct that few fold axes can be measured. At the exposures east of Libby slump folds occur higher in the section. One composite slump sheet at locality B shown on plate 2 is about 3 m thick, about 160 m above the base of the member, and extends at least 3.5 km northward along strike. The attitudes of the slump-fold axial planes in this sheet range considerably, but slumping was probably mostly northward.

Folded beds in the upper part of the upper member are exposed in a roadcut on the east side of Lake Koocanusa about 2 km northwest of the crossing of Tenmile Creek (pl. 2, loc. C). These may be slump folds, but cleavage along the fold axes suggests either a tectonic origin or tectonic modification. If these features are slump folds, the direction of slumping was to the southeast (pl. 2).

The contact of the upper member with the overlying transition member is rather sharp and, in adequate exposures, can generally be located to within 1 m. The contact is marked by the abrupt change from planar interlaminated argillite and siltite of the upper member to massive silty argillite and wavy interlaminated siltite and argillite of the basal part of the transition member. The nature of the contact is illustrated by the following stratigraphic section.

Measured Section 2

Valcour 7½-minute quadrangle. Roadcut along Montana Highway 37 on east side of Lake Koocanusa on the line between secs. 1 and 2, T. 34 N., R. 29 W. Measured by E.R. Cressman.

	Thickness (meters)
Prichard Formation (part):	
Transition member (basal part only):	
24. Siltite, medium-gray, argillic. Overlying beds not measured	1.4
23. Argillite and siltite: Silty argillite in beds about 0.1 m thick alternating with sets mostly 0.2–0.5 m thick but as much as 1 m thick of interlaminated argillite and siltite; laminae are wavy; some scour-and-fill structure and some cross-lamination present; cross-laminae dip both north and south. Unit contains some lenses as thick as 0.8 m of garnet- and hornblende-bearing white quartzite	6.3
22. Quartzite; light-gray, very fine grained; contains some faint, broadly wavy, continuous laminae; garnet- and hornblende-bearing white quartzite at base and near top; upper white quartzite pinches and swells	1.2
21. Argillite and siltite: Medium-light-gray biotite-bearing silty argillite in beds 0.1–1 m thick but mostly about 1 m thick that become more silty toward top of section, alternating with sets 0.05–0.8 m thick of medium-gray argillite and very light gray siltite in alternating beds about 3 cm thick; beds mostly planar, continuous, but wavy nonparallel in part; some cross-lamination and some ripple marks present. Unit contains some lenses of garnet- and hornblende-bearing white quartzite	9.5
20. Argillite and siltite interlaminated; laminae planar, continuous; graded in part; argillite laminae 1–3 mm thick, siltite laminae 0.1–1 mm thick	0.24
19. Argillite, light-gray, biotite-bearing	0.84
18. Argillite and siltite: Medium-gray biotite-bearing argillite in beds mostly 0.1 m thick alternating with sets about 0.2 m thick of interlaminated medium-gray argillite and very light gray siltite in laminae 0.5–1 cm thick; siltite laminae wavy continuous to lenticular and cross-laminated; cross-laminae dip to west. Contains some lenses of garnet- and hornblende-bearing white quartzite	5.0
17. Argillite and siltite; similar to unit 21	11.6
16. Quartzite, white, garnet- and hornblende-bearing	0.12
15. Argillite, light-gray, silty, biotite-bearing; planar continuous laminae at base, very irregular laminae at top. Unit is slump fold	0.5
Thickness of transition member measured	36.7

Prichard Formation (part):—Continued	*Thickness (meters)*
Upper member (upper part only):	
14. Argillite and siltite, biotite-bearing, interlaminated, very thinly interbedded: Medium-light-gray argillite in planar continuous laminae and beds 1 mm – 20 cm, averaging 4 cm, thick; very light gray siltite in planar continuous laminae and beds 1 mm – 6 cm, averaging 4 cm, thick; some siltite layers are lenticular and cross-laminated. Contains some graded siltite-argillite couplets 2–4 cm thick	0.5
13. Argillite, medium-light-gray; faint lamination indicates bed is slumped unit; base cuts across bed below; unit is lenticular	0–0.75
12. Argillite and siltite; similar to unit 14	3.25–4.0
11. Quartzite, white; contains garnet and hornblende; faint laminae suggest climbing ripples	0.1
10. Argillite and siltite, biotite-bearing, interlaminated, very thinly interbedded: Medium-light-gray argillite in planar continuous laminae and beds <1 mm – 4 cm, averaging 1.5 cm, thick; light-gray siltite in planar continuous laminae and beds 1 mm – 2 cm, averaging 5 mm, thick; a few siltite beds are lenticular and cross-laminated; contains a few graded siltite-argillite couplets	1.55
9. Argillite and siltite; similar to unit 10; contains lens of garnet- and hornblende-bearing white quartzite 5 cm thick and 0.9 m above base	1.65
8. Argillite; at top, dark-gray garnet-bearing siltite 2–6 cm thick	0.1
7. Argillite and siltite; similar to unit 10	1.74
6. Argillite; indistinct laminae suggest unit is slumped; contains lenses of garnet- and hornblende-bearing white quartzite	0.1
5. Argillite and siltite; similar to unit 10	0.2
4. Argillite, light-gray	0.13
3. Argillite and siltite; similar to unit 10; abundant fluid-escape structures at base where unit 2 is not present	0.16–0.3
2. Quartzite, white, garnet- and hornblende-bearing; about 2 m long	0–0.14
1. Argillite and siltite; similar to unit 10; concealed below	1.7
Thickness of upper member measured	11.18–12.96

TRANSITION MEMBER AND THE UPPER CONTACT OF THE PRICHARD FORMATION

Ransome and Calkins (1908, p. 29, 30) noted that near Prichard Creek the banded "blue gray" argillite of the Prichard grades upward into light-greenish-gray quartzite and argillite of the Burke Formation, and they placed the formational contact at the top of the highest "considerable body of blue-gray argillite." Thus, the transitional rocks were included in the Prichard. Subsequent workers have either followed the original definition by including the zone of transition in the Prichard or have mapped it with the overlying Burke Formation. Hobbs and others (1965, p. 33–34) in the Coeur d'Alene mining district locally mapped the zone separately and included it in the Prichard as an informally designated upper part. Similarly, Harrison and Jobin (1963, 1965) near Pend Oreille Lake mapped the zone as the argillite and siltite member of the Prichard Formation. In the Kalispell 1°×2° quadrangle the transition zone has been mapped everywhere except in the Cabinet Wilderness as the transition member of the Prichard Formation. Conversely, Wells (1974), mapping near Alberton, Mont., included the transitional rocks in the Burke Formation as did Wells and others (1981) in the Cabinet Mountains south of Libby. In the Plains-Perma area I placed the top of the Prichard at the top of member H for local convenience (Cressman, 1981; 1985, p. 38). On the geologic map of the Wallace 1°×2° quadrangle (Harrison and others, 1986) the contact with the Burke is shown in some areas at the base of the transition member and in other areas at the top. In Canada the transitional unit is placed in the Creston Formation that overlies the Aldridge (J.E. Reesor, oral commun., 1981). In this paper I follow the original usage of Ransome and Calkins (1908), and the section of mixed rock types that lie between planar-laminated argillite and siltite of member H or the upper member of the Prichard Formation below and blocky-weathering greenish-gray to yellowish-gray argillite and siltite of the Burke Formation above are assigned to the transition member of the Prichard.

The transition member is present wherever the upper contact of the Prichard Formation is exposed. Throughout much of the area the member is about 800 m thick, but significant variations from that figure can result from inconsistencies in locating the contact with the Burke Formation. For these reasons I do not present a map showing distribution and thickness of the member.

The transition member generally consists of three parts—a basal unit of blocky silty argillite, a middle unit of irregularly interlaminated siltite and argillite, and an upper unit similar to the middle but containing interbeds of quartzite. The relative thicknesses of the units differ from place to place, but the units can be identified nearly everywhere. The contacts between these parts are gradational, commonly through several tens of meters, but they are mappable, at least locally, at a scale of 1:24,000 (M.G. Lis, oral commun., 1980).

The lower massive part of the transition member consists of argillite or silty argillite in beds mostly about 0.1–0.5 m thick that alternate with thinner sets of interlaminated siltite and argillite in which the laminae range from planar to wavy (fig. 26). A few quartzite beds may also be present. Locally, slump-folded units, some as thick as 3 m, are present. These basal beds are described in Measured Section 2. I have not seen the garnet- and hornblende-bearing white quartzite lenses that are described elsewhere in the transition member. This basal part is commonly only a few meters to a few tens of meters thick, but in the outcrop area between Libby and Kalispell it makes up about one-third of the total transition member.

The middle part of the member consists of sets of interlaminated light-gray siltite and dark-gray argillite that alternate with beds and sets of light-gray to yellowish-gray siltite (fig. 27). The thickness of the sets and their relative abundance vary considerably from place to place. In the interlaminated sets siltite and argillite are present in about equal amounts. The laminae are mostly continuous and nearly planar to broadly wavy and irregular, though some of the siltite

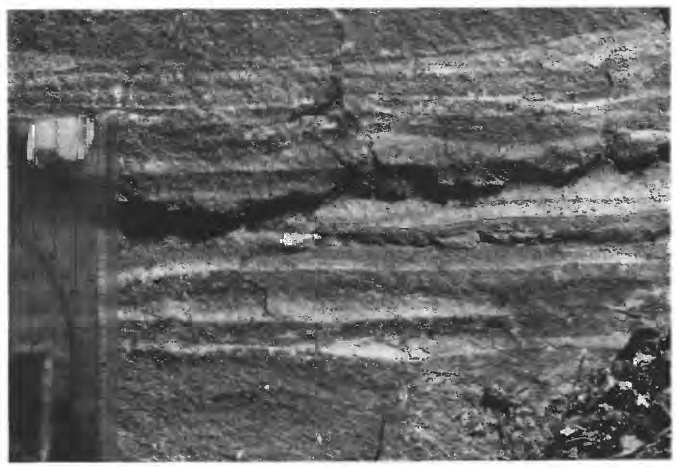


FIGURE 26.—Interlaminated siltite and argillite in basal part of transition member of Prichard Formation. Siltite laminae are light colored.

FIGURE 27.—Exposure in middle part of transition member of Prichard Formation. Sets of interlaminated siltite and argillite alternate with siltite beds (s). Note broadly wavy nature of laminae.

layers are lenticular. Small-scale scour-and-fill structure, cross-lamination, and fluid-escape structure may also be present. Discontinuous shrinkage cracks are common in the argillite layers; they consist of two or three short cracks that radiate from a common center at angles of 90°–120°. The cracks, which do not join to form polygonal networks, have served to some extent as passages for fluid escape. Inasmuch as there is no other evidence of subaerial exposure and inasmuch as the cracks resemble those produced subaqueously by an increase in the salinity of the water (Burst, 1965), these features were probably produced by syneresis rather than by desiccation.

The third and uppermost part of the transition member is commonly 100–150 m thick and consists of quartzite, siltite, and interlaminated siltite and argillite. The interlaminated siltite and argillite is like that of the middle part of the member and is in sets that average about 0.2–0.4 m thick. The siltite is in even beds mostly 0.1–0.3 m thick and ranges in color from medium gray to yellowish gray and greenish gray; the yellowish and greenish hues become more prominent upsection. The quartzite is fine to very fine grained, poorly sorted, argillaceous, and in even beds mostly 0.05–0.3 m thick; the quartzite beds are mostly structureless, but some contain planar to wavy, continuous laminae. Biotite porphyroblasts become conspicuous upsection as the rocks become lighter in color. Some beds contain molds of pyrite cubes as much as 2.5 cm across.

Calcareous siltite is present in the transition member from Glacier National Park as far west as the exposures west of Troy, Mont., and as far south as near Plains. Except in the Park these calcareous beds are mostly in the upper part of the member and make up only a small part of the total interval. K.D. Loos (written commun., 1985) determined the carbon isotope ratios ($\delta^{13}C$) to range from -17.10 to -17.61. This suggests that the carbonate was derived from organic matter during sulfate reduction by anaerobic bacteria.

Ripple marks are widespread in quartzite and siltites of the transition member, but they are not abundant. Most occur in the upper part. The ripples commonly have wavelengths of 6–10 cm and amplitudes of 6–10 mm. Crest lines are commonly sinuous, and many are short-crested current ripples (fig. 28). Hexagonal interference ripples are exposed at one locality (fig. 29).

The irregular laminae, the abundance of scour-and-fill structure, and the presence of both current and wave ripples all suggest that most of the transition member was deposited from or reworked by traction currents. K.D. Loos (written commun., 1985) measured paleocurrent directions at three locations east-northeast of Libby; the directions range from east-northeast to west-southwest, but the mode, which is well defined, is about N. 45° W.

FIGURE 28.—Short-crested ripple marks on siltite or very fine grained quartzite of transition member of Prichard Formation.

FIGURE 29.—Interference ripple marks on siltite or very fine grained quartzite of transition member of Prichard Formation. Exposed in cut on Vermillion Creek road 32 km N. 20° W. of Thompson Falls, Mont.

The contact of the transition member of the Prichard Formation with the overlying Burke Formation has been very difficult to locate consistently. As noted above, Ransome and Calkins (1908, p. 29, 30) placed the contact at the highest "considerable body of blue-gray argillite." This definition is difficult to follow because "considerable" is imprecise, because the color changes gradually upsection from neutral hues to yellowish and greenish gray, and because Ransome and Calkins did not describe the sedimentary structures that now are so important in identifying the units. Because of the difficulty in applying the original definition, some workers, noting that iron in the Prichard occurs mostly as a sulfide whereas in much of the Burke it is in magnetite, have placed the contact at the change from iron sulfides below to magnetite above. This definition has the advantages that the contact can be located closely and the results are repeatable, and I have used it locally in mapping. However, the definition presents difficulties. First, the boundary between sulfide-bearing rocks below and magnetite-bearing rocks above is a metamorphic contact that is almost certainly at the same position as an original diagenetic contact between sulfides and oxides. As such, the contact may vary considerably from lithologic contacts that reflect changes in the depositional environment. Second, in several localities intervals containing sulfide alternate with zones containing magnetite. Third, at many localities the use of this definition includes rocks in the Prichard Formation that by all other criteria would be placed in the Burke. Therefore, in this paper the Prichard-Burke contact is placed at the highest unit of laminated light-gray and dark-gray siltite and argillite that exhibits small-scale sedimentary structures such as scour-and-fill and cross-lamination and weathers a rusty-brown color. This is the definition that was used in mapping most of the Kalispell 1°×2° quadrangle.

Measured Section 3, measured on the west side of Lake Koocanusa in roadcuts along Montana Highway 37 on the line between sections 1 and 2, T. 34 N., R. 29 W., illustrates the nature of the formational contact.

Measured Section 3

Inch Mountain 7½-minute quadrangle. Roadcuts along Montana Highway 37 on west side of Lake Koocanusa on the line between secs. 1 and 2, T. 34 N., R. 29 W. Measured by E.R. Cressman

	Thickness (meters)
Prichard Formation (part):	
Burke Formation (basal part only):	
7. Siltite and argillite: Light-gray siltite in beds 0.05–0.2 m thick alternates with sets 0.08–0.15 m thick of very thinly interbedded medium-gray argillite and dark-gray siltite in continuous wavy nonparallel beds about 5 cm thick. Minute magnetite octahedra dispersed through rock. Weathers yellowish gray to grayish yellow. Only 15.7 m measured, but overlying rocks are similar	15.7
Prichard Formation (part):—Continued	
Burke Formation (basal part only):—Continued	
6. Similar to unit 7, but magnetite is absent and oxidized pyrite cubes are scattered through unit	3.7
5. Argillite, yellowish-gray, silty, laminated; laminae mostly wavy continuous; contains a few unlaminated intervals about 5 cm thick and some fluid-escape structures	2.5
4. Covered	18.3
3. Argillite and siltite, interbedded: Medium-gray argillite in beds mostly about 0.15 m thick alternates with medium-light-gray biotite-bearing siltite in beds of about the same thickness; beds are planar to broadly wavy; contacts between beds mostly gradational, but load casts present at base of some siltite beds; oxidized pyrite cubes present in some layers; argillite weathers light gray, siltite medium gray. Grades through 1 m from unit below	5.2
Thickness of Burke Formation measured	45.4
Transition member of Prichard Formation (upper part only):	
2. Argillite and siltite, interlaminated and very thinly interbedded: Medium-gray argillite in laminae and beds 1 mm – 5 cm thick; light-gray to light-greenish-gray argillic siltite in irregular continuous and discontinuous beds 1–2 cm thick; siltite contains some small-scale scour-and-fill structure and some ripples with an amplitude of 1.5 cm and a wavelength of 20 cm. Unit weathers moderate brown to yellowish brown. Grades through 2 m from unit below	9.8
1. Argillite, siltite, and quartzite, interlaminated and interbedded: Interlaminated and very thinly interbedded argillite and siltite in sets 0.1–0.9 m thick alternate with quartzite in beds 0.05–0.2 m thick; argillite is medium gray, in laminae and beds mostly 1–2 cm thick; siltite in laminae mostly 0.05–1 cm thick; laminae mostly wavy, nonparallel; many siltite layers pinch and swell; some scour-and-fill structures about 1 cm thick, some cross-lamination, and some fluid-escape structures; quartzite is very light gray, fine grained, and biotite bearing; quartzite mostly structureless, but contains some wavy, continuous laminae, and ball-and-pillow structure present at base of some beds; most quartzite beds have gradational upper and lower contacts. Exposure weathers moderate brown to yellowish brown	12.2
Thickness of transition member of Prichard Formation measured	22.0

In Measured Section 3 the differences between the transition member of the Prichard Formation and the lower part of the Burke Formation are chiefly in the thickness and character of the sedimentary layers and in the weathering colors. In the transition member the sedimentary layers are mostly less than 5 cm thick, whereas in the Burke they are mostly 5 cm or more thick. In the transition member the layers mostly have sharp upper and lower contacts and exhibit small-scale scour-and-fill and cross-lamination, whereas in the lower Burke contacts between argillite and siltite are commonly gradational, and small-scale sedimentary structures are absent. Finally, the transition member weathers moderate brown to yellowish brown, whereas the basal Burke weathers yellowish gray to grayish yellow. Note that in Measured Section 3 the change from pyrite below to magnetite above occurs within an otherwise uniform lithologic unit and well above the contact selected by the criteria described above.

Near McDonald Creek on the west side of Glacier National Park 350 m of section between the planar-laminated and very thinly interbedded argillite and siltite of member H of the Prichard Formation and green argillite and siltite typical of the Appekunny Formation have been called the Wolf Gun Member of the Appekunny Formation by Earhart and others (1983). The Wolf Gun Member is in approximately the same stratigraphic position as the transition member of the Prichard Formation as described in this report. The Wolf Gun Member at its type locality near Anaconda Creek, Mont., about 13 km northwest of McDonald Creek consists of three parts (Earhart and others, 1983). These are (1) a lower unit of black and greenish-gray siltite and argillite, calcareous in part, with wavy, discontinuous, nonparallel laminae; (2) a middle unit of limestone, calcareous siltite, and stromatolitic zones; and (3) an upper unit of thin-bedded green, dark-greenish-gray, and tan argillite. Mud cracks, fluid-escape structures, and mud-chip breccias are common. Though in a similar stratigraphic position, the Wolf Gun of the Anaconda Creek area differs from the transition member of the Prichard Formation in its green color, the abundance of carbonate, the discontinuous nature of the laminae, and the presence of mud-chip breccia.

Whipple and Harrison (1987) mapped two lithologic units together as the upper part of the Prichard Formation in the Whitefish Range. My reconnaissance indicates that the uppermost of the two units is very similar to the transition member as exposed nearly everywhere else in the area of this report; it consists of the same three parts that have been described as making up the transition member elsewhere, and the uppermost Prichard there differs from the transition member to the west mainly in containing slightly more calcareous siltite. The transition member in the Whitefish Range is overlain by a unit of uncertain affinity that consists of olive-colored gray and buff siltite and very fine grained quartzite that weather brown and contain abundant specks of siderite (J.E. Harrison, oral commun., 1986). This unit is overlain by the Spokane Formation.

Inasmuch as the Wolf Gun Member of Glacier National Park and the uppermost Prichard in the Whitefish Range are so similar to the transition member of the Prichard Formation throughout most of the rest of the report area, the name Wolf Gun could, with some logic, be applied to all of the transition member. However, I do not think that a name derived from near the outer limits of occurrence of the unit and from an atypical section should be applied to the unit. Furthermore, the relation of the top of the Wolf Gun in Glacier National Park, and to a lesser extent in the Whitefish Range, to the top of the transition member elsewhere in the area is by no means clear. Therefore, in this report I do not use the name Wolf Gun and instead refer the entire unit, including the section in Glacier National Park, to the transition member.

LITHOCORRELATIONS

Lithocorrelations of the stratigraphic units described above are shown on plate 4. Most of the lithocorrelations are obvious and unambiguous, but a few uncertainties exist. The most serious uncertainties concern correlations of the basic sills. Seismic lines indicate that sills or zones of sills are continuous over great distances, and some individual sills are known to be of considerable extent. For example, the highest sill, shown in section 1, plate 4, undoubtedly extends at least 80 km, and nearby the next sill down (shown on pl. 4) certainly extends the same distance. However, some of the sills are shown as being continuous for as much as 200 km. This is an extraordinarily long distance, and any such correlation is surely hypothetical. Furthermore, an examination of the correlation of the sills, for example between sections 2, 9, and 11 on plate 4, will show that other plausible correlations are possible.

Another questionable correlation is that of the argillite member of the quartzite facies with the argillite facies. On plate 4, sections 10 and 11, the argillite member is shown as a tongue of the lower part of member H, but another possibility is that the argillite member is an extension of member F. The first correlation, that shown on plate 4, is more in accord with thickness trends of the quartzite member.

Although plate 4 presents lithocorrelations, some time relations can be inferred from the lithologic relations as will be discussed in a subsequent part of this report.

AGE

The maximum possible age of the Prichard Formation is that of the pre-Belt crystalline basement of Montana and Alberta. The major metamorphic event that affected the pre-Belt basement of southwest Montana was at 3,730±85 Ma (James and Hedge, 1980), but the isotopic clocks were reset by a thermal event at

1,600±100 Ma (Giletti, 1966). In Alberta, K–Ar dates on biotite indicate that the basement was consolidated between 1,650 and 1,850 Ma (Burwash and others, 1962), and Evans and Fischer (1986) reported a U–Pb zircon age of 1,576±13 Ma from the Priest River Group near Spokane, Wash. In the northern part of the Idaho batholith U–Pb ages of xenocrystic zircon imply basement ages of 1,625–2,349 Ma (Toth and others, in press; Bickford and others, 1981). Thus the maximum age of the Prichard is about 1,600 Ma. The minimum age of the Prichard is 1,330±45 Ma, a K–Ar date obtained from metamorphic biotite in the upper part of the Prichard near Alberton, Mont. (Obradovich and Peterman, 1968, p. 745).

One other applicable date has been published, a 1,433±30 Ma U–Pb zircon age for the Crossport C sill (Zartman and others, 1982). The Crossport C sill is identified on section 10, plate 4, where its identification is based on correlation from section 11A. In order to know what this dates, other than the sill itself, one must know the depth at which the sill was intruded. I am unaware of any evidence by which the depth of intrusion of the Crossport C sill can be determined directly, but consideration of sills elsewhere in the Belt, of the mechanics of sill-intrusion, and of the depth of intrusion of sills in rocks other than the Belt all provide some constraints.

Direct evidence in the Prichard Formation of the depth of sill intrusion is scanty. Near Perma, Mont., and immediately east of Plains, a diorite sill 300 m thick rises about 600 m in the section (Cressman, 1985, p. 43). Obviously, the deeper part of the sill was intruded under more than 600 m of cover. Other evidence in the same area indicates that sills were intruded under a considerable range of thicknesses. Some sills are underlain and overlain by zones of hornfels as much as 280 m thick (R.L. Earhart and R.E. Van Loenen, written commun., 1980), whereas the hornfels zones of other sills in the same area are too thin to map at a scale of 1:24,000. One well-exposed sill near Sylvanite, Mont., is 60 m thick but has no hornfels at the upper contact, and the overlying beds are deformed through a thickness of only 0.5 m (fig. 30). The sills accompanied by thick zones of hornfels were probably intruded at greater depths where the ambient temperature was higher and the content of pore water less than were the sills with thin or no hornfels.

The maximum depth of intrusion of one sill in the Prichard Formation can be inferred from the maximum depth at which fluidization of the sediment was likely to have occurred. The massive member of the Prichard is interpreted to have resulted from fluidization as a consequence of sill intrusion where heat from the sill converted pore water to steam and increased the pore pressure. According to Kokelaar (1982), fluidization by

FIGURE 30.—Upper contact of diorite sill with argillite and siltite of quartzite member of Prichard Formation. Hammer head is on contact. Exposed in cut along Yaak River road.

such a process is unlikely at depths where the pressure is much greater than that of the critical point of water, which Kokelaar gives as 312 bars for water with a salinity of 34.5 parts per thousand (assuming all the solutes are NaCl). Lithostatic pressure is given by $P=\rho dg$ where P is lithostatic pressure, ρ is the average density of the overburden, d is the depth of overburden, and g is the gravitational constant. Assuming an average density of 2.74 g/cm^3 for the Prichard Formation (Harrison and others, 1972, table 3), a pressure of 312 bars would be attained at a depth of 118 m in terms of the present thickness of the section. Assuming an average void ratio at that time of 0.6, the depth below the sediment surface at the time of intrusion would have been nearly 200 m. Inasmuch as the massive member is 150–200 m thick and inasmuch as the calculation does not consider water depth, the calculation can be taken only as showing that the sill was intruded at very shallow depth.

Two hypotheses have been proposed to explain sill intrusion. These are the hydrostatic hypothesis proposed by Gilbert in 1880 and the tectonic hypothesis of Anderson (1951, p. 50–53, 153–155). According to the hydrostatic hypothesis, sill intrusion will occur where the magma in its upward passage encounters rocks of equal or lesser density than itself. The density of basaltic magma was given by Roberts (1970) as 2.98 g/cm^3, whereas the density of rocks of the Prichard Formation in their present compacted and metamorphosed state is 2.74 g/cm^3 and would have been less at the time of intrusion. Therefore, according to the hydrostatic hypothesis, sill intrusion could have occurred at any depth in the Prichard to which pressure in the magma chamber could lift the magma.

Einsele (1982) in his study of sills that intrude sediments of Pliocene and Quaternary age in the Guayanas

Basin in the Gulf of California noted that tensile strength must be considered. For sill intrusion to occur the magma pressure must exceed the combined lithostatic pressure and tensile strength of the country rock. In the Gulf Coast the consolidation of argillaceous sediments seems to proceed gradually and steadily with burial, but at a depth of about 1,800–2,400 m late-stage dehydration involving the montmorillonite-illite transition begins (Perry and Hower, 1972). Although no data on tensile strength at these depths are available, it is at least possible that it might increase enough so that the combined tensile strength and lithostatic pressure might exceed the magma pressure. If so, intrusion at greater depths would be unlikely. Thicknesses of 1,800–2,400 m are for the sediments at the time of intrusion. Assuming an average void ratio at that time of 0.6, these thicknesses would be the equivalent of 1,125–1,500 m of completely consolidated rock.

The tectonic hypothesis states that sill intrusion will occur where the minimum principal stress, which is horizontal at greater depths, becomes vertical at shallower depths. The maximum depth of sill-intrusion can be expressed by the following equation (Roberts, 1970, p. 329):

$$Z=vh/(1-n)(1-v)$$

where
Z=maximum depth of sill intrusion,
v=Poisson's ratio,
h=hydrostatic head of magma at the earth's surface, and
$n=v/(1-v)$.

Roberts (1970) assigned a value of 2–4 km to h, based on the heights of basaltic volcanoes. Poisson's ratio for shale and quartzite ranges widely but averages about 0.15 (Birch, 1966, p. 167). Using these values, Z ranges from 0.5 to 1 km.

Mudge (1968) compiled data from 54 localities, mostly in the western United States but none in the Belt basin, and found that most sills intruded beneath a cover of 900–2,300 m. However, Deep Sea Drilling Project (DSDP) cores in the Gulf of California have encountered sills at depths as little as 50 m (Curray and others, 1982).

In summary, one sill in the Prichard Formation was intruded under at least 0.6 km of cover, and fluidization of sediment above another suggests intrusion under less than 0.12 km of sediment. The hydrostatic hypothesis suggests intrusion at less than 1.5 km, and the tectonic hypothesis suggests intrusion depths of less than 1 km. All these depths are in terms of the present thickness. Elsewhere in North America, most sills intruded beneath 0.9–2.3 km of cover, but they can intrude at depths as shallow as 50 m. Although the evidence is far from conclusive, the sills, including the Crossport C, were probably intruded under no more than 1.5 km and probably less than 1 km of cover in terms of the present thickness. Therefore, the 1.433±30 Ma age of the Crossport C sill probably applies to rocks somewhere within the bracket shown on section 10, plate 4. Thus, about half the Prichard in that area is older than 1,433 Ma and half is younger.

Schmidt and Rowley (1986, p. 283) stated that where good quality analyses are available, the age range of the principal phase of basaltic magmatism in continental rifting tends to be only 10–15 m.y. Assuming this to apply to the sills of the Prichard Formation and assuming the maximum depth of sill intrusion to be 1 km, a maximum age of 1,448 Ma would apply to an horizon 1,100 m above the base of the section; similarly the base of the transition member, about 1 km above the highest sill, would be no younger than 1,418 Ma. Thus, about 70 percent of section 10, plate 4, would have been deposited through an age range of no more than 30 m.y. However, there is some evidence for at least two periods of sill intrusion into the Belt Supergroup (Harrison and others, 1974, p. 5).

Many studies of sedimentation on passive continental margins have shown that in those environments tectonic subsidence is directly proportional to the square root of time. The procedure for determining tectonic subsidence is called backstripping and consists of measuring the thicknesses of successive intervals of well-dated sediments whose depth of deposition is known. That part of the subsidence that cannot have been caused by loading is the tectonic subsidence. In the following discussion the procedure is reversed, and I will attempt to determine time from the subsidence curve, which is assumed to be a straight-line function of the square root of time. The subsidence curve will be calibrated by two dates. The first and most reliable is the 1,433-Ma date of the Crossport C sill, which is taken to date rock 600 m above the sill. This position is near the middle of the range inferred for the depth of intrusion. The other date is 1,375 Ma for the base of the Missoula Group based on a comparison of the polar wandering curve and polarity zoning of the Belt Supergroup with paleomagnetic data from more closely dated Middle Proterozoic rocks elsewhere (Elston and Bressler, 1980).

The section analyzed (fig. 31) is from the north side of Pend Oreille Lake and is taken from plate 2 of Harrison and Jobin (1963) except for the Prichard Formation, which is from stratigraphic section 10 of this report. The base of the Missoula Group correlates in this section with the base of an argillite, siltite, and limestone member of the Wallace Formation (J.E. Harrison, oral commun., 1986), and the 1,374-Ma date of Elston and Bressler (1980) is so placed on the column in figure 31*A*. The depth of water shown in figure 31*A* is based on the presence of shallow-water features in most of the Belt Supergroup above the Prichard Formation, on the identification of the quartzite member as a submarine fan, and on the observation that most large modern fans extend to depths of 2,000–3,000 m (see papers in Bouma and others, 1985a).

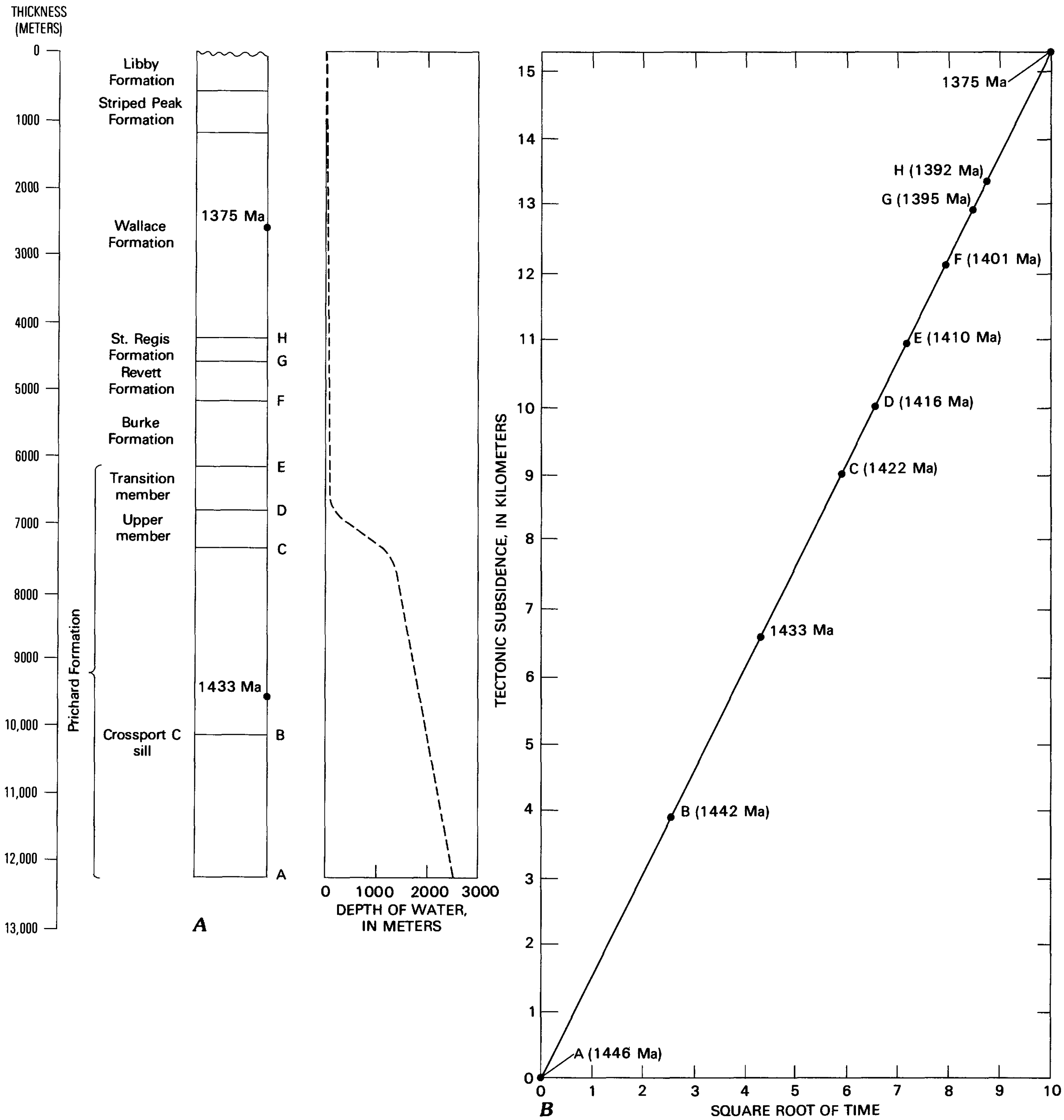


FIGURE 31.—*A*, Stratigraphic section on north side of Pend Oreille Lake showing ages used in subsidence analysis and inferred depth of water during deposition. *B*, Relation between tectonic subsidence and square root of time. Time to deposit section from base to 1,375 Ma considered 100 percent. Dates in parentheses are inferred from curve.

From Steckler and Walls (1978, p. 86), the general equation for backstripping is

$$Y=\phi^*S^*\left[\frac{(\rho_m-\rho_s)}{(\rho_m-\rho_w)}-SL\frac{\rho_w}{(\rho_m-\rho_w)}\right]+Wd-SL$$

where

Y=depth of basement without the effects of sediment and water loads,
ϕ^*=basement response functions,
S^*=sediment thickness corrected for compaction,
ρ_m=mean density of mantle,
ρ_s=mean density of sediments,
ρ_w=mean density of water,
SL=sea level relative to present day, and
Wd=water depth.

The following assumptions are made for Belt time:

$\phi^*=1$ (the value for an Airy-type basement),
$\rho_m=3.33$ g/cm^3 (Watts, 1981, table 4),
$\rho_w=1.03$ g/cm^3 (Watts, 1981, table 4), and
$SL=0$ (this is unlikely but the only choice possible because of lack of evidence).

The values of S^* and ρ_s were determined from the present thicknesses of the units shown in figure 31*A*, the density of these units given in Harrison and others (1972, table 3), and the averaged porosity-depth curve in Watts (1981, fig. 8). The value of Y was then calculated for the intervals on figure 31*A* of A–B, B - 1,433 Ma, 1,433 Ma - C, C–D, D–E, E–F, F–G, G–H, and H - 1,375 Ma, and used to locate the various points on the vertical axis of figure 31*B*. Tectonic subsidence is assumed to be a linear function of the square root of time, and time is expressed as percent of the total time to deposit the section from A to the position of the 1,375–Ma date. Having two dated horizons, that of the 1,375–Ma date and that of the 1,433–Ma date, it is then possible to calculate the dates shown in parentheses in figure 31*B*.

This analysis indicates that the Prichard Formation in this section was deposited through an interval of 36 Ma, beginning at 1,446 Ma and ending at 1,410 Ma. The transition member represents 6 Ma, the upper member 6 Ma, and the rest of the formation 24 Ma. The accumulation rate of the quartzite member, 40 cm/10^3 yr that was calculated from these dates is reasonable for the inferred environment. Accumulation rates for the rest of the Belt would have been much slower. For example, the rate for the Wallace Formation below the position of the 1,375–Ma date would have been 0.18 cm/10^3 yr, corrected for compaction.

Members A–E of the argillite facies of the Prichard Formation are probably older than any of the rocks exposed in the Pend Oreille Lake area. Therefore, if the above analysis is correct, deposition of the Prichard must have begun before 1,446 Ma.

The validity of the above analysis depends on assumptions that the two assigned dates are correct and that the Belt Supergroup resulted from one major episode of subsidence and deposition; both assumptions have been questioned. The 1,433–Ma date on the Crossport C sill is reliable, but as discussed previously, the exact position in the section to which the date applies is not as certain. The other date, 1,375 Ma for the base of the Missoula Group as inferred by Elston and Bressler (1980) from the polar-wandering curve, is in conflict with radiometric ages obtained by Obradovich and Peterman (1968), and the validity of dates assigned to the polar-wandering curve has been seriously questioned by Obradovich and others (1984). The assumption that the Belt resulted from a single major episode of subsidence and sedimentation is in conflict with radiometric ages that suggest two major hiatuses, each of 200-m.y. duration. Thus the determination of ages from the subsidence curve is highly speculative.

In summary, the Prichard Formation is younger than 1,576 Ma, the pre-Belt basement, and older than 1,330 Ma, the age of metamorphic biotite near the top of the formation. The middle part of the quartzite facies is dated at 1,433 Ma by the age of the Crossport C sill, and speculative subsidence studies suggest that the quartzite facies may have been deposited between 1,446 Ma and 1,410 Ma. That part of the argillite facies below the top of member E would then be dated as older than 1,446 Ma and younger than 1,576 Ma. An analysis based on the assumption that the sills were intruded in a single episode 15 m.y. long also suggests that the quartzite facies was deposited in a few tens of million years.

DEPOSITIONAL ENVIRONMENTS AND PALEOGEOGRAPHY

THE ENVIRONMENT IN MIDDLE PROTEROZOIC TIME

Analogy with present-day events and processes is the basis of all interpretation in historical geology, but the older the rocks the more the caution that must be used in making direct comparisons. In Middle Proterozoic time not only was the biota vastly different from that of today, but the physical environment differed in important aspects from that of the present. Although knowledge of those times is fragmentary, the following paragraphs will examine what is known and what can be inferred as to how environmental conditions in the Middle Proterozoic differed from those of today.

Because of tidal friction the rotational velocity of the earth has slowed through geologic time. The amount of slowing that is inferred depends on which model of the earth-moon system is accepted, but a reasonable estimate is that at 1.5 Ga the day was about six hours shorter and the rotation rate about 1⅓ times faster than today (Walker and others, 1983, fig. 11-1), though projection of rates derived from paleontologic evidence in the Phanerozoic suggests a rotation rate 2–2½ times that of today (Hunt, 1979). The consequences of a faster rate of rotation would have been a stronger Coriolis force, reduced wind speeds, a shallower oceanic mixed layer, less precipitation, and colder polar regions (Hunt, 1979).

The oxygen content of both the atmosphere and the hydrosphere has increased during geologic time as a consequence of oxygen-producing photosynthesis. In Middle Proterozoic time the oxygen content of the atmosphere must have been at least 0.01 PAL (present atmospheric level) or 2×10^{-3} to 4×10^{-3} atmosphere, which is the minimum amount needed to support aerobiosis (Chapman and Schopf, 1983). The maximum oxygen content was probably 0.02 PAL, a value consistent with data on paleosols (Holland and Zbinden, 1985). Other evidence suggests about the same oxygen level. Rhoads and Morse (1971) noted that on the sea floor of modern oxygen-deficient basins the change from anaerobic conditions with no metazoans to dysaerobic conditions with a soft-bodied infauna occurs at a dissolved oxygen content of 0.1–0.3 ml/l, which corresponds to about 0.01 PAL. They suggested that this was the oxygen level at which metazoans made their first appearance in Late Proterozoic time.

The best evidence, then, indicates that the oxygen content of the Middle Proterozoic atmosphere was only 1–2 percent of that of today. The oxygen content of the oceanic mixed layer would have been correspondingly lower, and the mixed layer would have been thinner than it is today because of the effects of faster rotation of the planet. Wilde (1987, p. 454) has written "Due to the oxygen demand of organic matter derived from photosynthetic organisms***any oxygen concentration in the atmosphere below about 5 percent PAL would be insufficient to maintain oxic conditions in the water column except in the uppermost few meters." The oxygen content of the ocean beneath the mixed layer would also have been less, and the oxidation of organic matter should have created anaerobic conditions over very large areas, especially inasmuch as organic productivity of the oceans was about the same then as it is now (see below). Wilde's (1987, fig. 1) calculations indicate that for nonglacial conditions and atmospheric oxygen at 1 PAL, the oceans would have remained anoxic to a depth of 3,000 m, though Wilde has cautioned that his results are approximate at best.

As the oxygen content increased through time, the CO_2 content decreased, also largely as a result of photosynthesis. Des Marais (1985) estimated that at 3 Ga the atmosphere contained at least two orders of magnitude more CO_2 than it does today. The CO_2 partial pressure of the atmosphere in Middle Proterozoic time must also have been appreciably greater than it is today (though see Sandberg, 1985), and the oceans were saturated or supersaturated with respect to calcite and dolomite (Holland, 1984, p. 422). One result of the higher CO_2 partial pressure was that chemical weathering and soil-forming processes, which should have been slower in the absence of a plant cover, were not much different from present-day rates (Cawley and others, 1969; Gay and Grandstaff, 1980; Holland, 1972, p. 648; Schau and Henderson, 1983).

In the absence of silica-secreting organisms the SiO_2 content of the oceans must have been greater than today; Holland (1984, p. 442) estimated the concentration to have been between 20 and 120 ppm. The silica content may have been governed by reactions with clays or zeolites or by the solubility of amorphous silica. If it were the last, evaporite basins should have been the favored environment for silica extraction.

The biota in Middle Proterozoic time was almost certainly entirely microbial and included both sulfate-reducing and oxygen-producing organisms (Schidlowski and others, 1983, p. 173). Although some authors have reported evidence of bioturbation in Middle Proterozoic rocks, there is no convincing evidence of metazoans in rocks older than 950 Ma (Glaessner, 1983, p. 330). Rocks of the Prichard have not been bioturbated and contain neither body nor trace fossils.

Organic productivity of the oceans was probably not much different in the Middle Proterozoic than it is today. According to Schidlowski (1983, p. 215), "It may reasonably be assumed that the early Earth was populated by prokaryotic ecosystems of both benthic and planktonic type whose sizeable rates of productivity suggest that photosynthesis may have gained little in quantitative importance during subsequent evolution." Schidlowski's conclusion is based largely on the amount of organic carbon in ancient rocks and on considerations of carbon-isotope mass balance. Veizer (1983, p. 291) also concluded on consideration of isotope balance calculations that the organic carbon reservoir in most of Precambrian time could not have been much smaller than it is today.

Microorganisms may have inhabited the soil (Beeunas and Knauth, 1985), but none of the rooted plants that have stabilized soils from Devonian time on had yet evolved. As a result, denudation and runoff rates should have been high and floods large (Schumm, 1968), but this may not have been the case because, as a consequence of the earth's faster rotation, precipitation was less.

MINOR-ELEMENT CONTENT AND THE ENVIRONMENT

Fifty-nine samples from all members of the argillite facies between Plains and Perma except the C and transition members were analyzed semiquantitatively by spectrographic methods. The results are listed in table 5.

TABLE 5.—*Semiquantitative spectrographic analyses of samples from the Prichard Formation near Plains, Montana*

[Numbers in parentheses below elements indicate lower limits of detection: L, detected but below limit of determination; N, not detected. Zn analyzed by atomic adsorption. Other elements looked for but not found, except as noted, at their respective sensitivity limits include Ag (0.5), As (200), B (10), Nb (20), Sb (200), Sn (10), Te (2,000), U (500), W (100), and Zn (200). Spectrographic analyses by M.J. Malcolm, N. Conklin, and I. Hamilton; atomic adsorption analyses by P. Briggs and J. Crock]

Unit		Percent				Parts per million														
	Sample No.	Fe (.05)	Mg (.02)	Ca (.05)	Ti (.002)	Mn (10)	B (10)	Ba (20)	Be (1)	Co (5)	Cr (10)	Cu (5)	La (20)	Ni (5)	Pb (10)	Sc (5)	Sr (100)	V (10)	Y (10)	Zr (10)
Member H																				
Argillite	9001	3	1	L	0.3	300	50	700	3	5	70	10	50	7	70	20	100	150	30	200
	9002	3	1	L	.2	200	50	500	1.5	5	50	15	100	5	50	15	N	150	50	200
	9003	2	.7	.07	.5	300	50	700	2	N	70	15	30	5	30	20	150	150	15	200
	9004	2	.7	.05	.3	300	50	500	2	7	50	15	100	10	30	15	N	100	50	200
	[1]9005	3	.7	.07	.3	300	50	700	1.5	20	70	20	70	50	20	20	100	200	50	200
	9036	3	.7	L	.3	300	70	500	2	5	70	10	50	15	50	20	100	150	30	150
	9037	5	1	.05	.3	200	70	500	1.5	30	100	50	50	30	70	20	N	200	50	150
	9056	3	.7	L	.3	150	50	700	2	10	70	30	50	15	30	20	N	150	70	200
	9058	3	1	.05	.3	200	70	500	1.5	15	50	30	50	30	70	15	N	100	30	150
Siltite	9006	2	.7	.15	.2	300	50	300	1.5	10	30	10	30	20	20	10	150	100	30	200
	9007	3	.7	.05	.3	300	70	100	1.5	10	70	L	50	30	10	15	N	150	30	200
	9008	3	.7	.2	.2	500	70	700	2	10	70	10	70	20	30	15	L	100	70	200
Quartzite	9057	3	.3	.2	.2	200	30	300	1	10	20	10	50	20	20	15	150	100	20	200
Member G																				
Argillite	9028	3	.7	.05	.3	200	50	700	1.5	15	70	20	50	20	20	15	100	100	20	150
	9029	5	1	.05	.5	300	70	1000	2	20	100	30	20	30	30	20	100	150	30	150
Siltite	9024	1	3	.2	.3	300	30	300	1.5	5	30	10	50	10	50	10	100	70	30	300
Quartzite	9025	1.5	.3	.1	.2	200	50	300	1.5	5	30	5	50	10	50	5	100	70	20	300
	9026	1.5	.2	.05	.15	200	30	300	1	5	30	10	50	10	30	5	100	70	20	200
	9027	1.5	.3	.7	.3	500	50	300	1.5	5	50	7	50	10	50	7	100	100	30	300
	9030	1.5	.3	.2	.3	150	30	500	1	10	70	10	30	10	20	10	150	70	30	300
	9031	1.5	.3	.15	.15	200	30	300	1	5	20	15	30	10	30	7	100	50	20	200
Member F																				
Argillite	[2]9032	2	.7	L	.3	150	150	500	1.5	5	70	20	N	5	20	15	L	150	20	200
	[3]9033	2	1.5	.7	.3	300	70	500	1.5	20	100	50	50	30	70	15	200	150	30	200
	9034	3	1	.1	.5	500	70	700	1.5	10	150	10	20	15	20	30	N	200	50	200
	9035	5	1	.7	.3	700	70	300	1.5	20	70	50	50	30	70	15	150	150	30	150
	9038	3	1	.1	.5	200	100	700	1	10	70	10	30	10	15	15	N	150	50	200
	[4]9039	3	1	.7	.3	300	100	500	1.5	15	70	30	70	30	30	15	100	150	30	200
	9040	3	1	.5	.5	500	100	700	1.5	10	100	10	30	5	20	20	100	150	30	200
	[5]9044	3	1	1	.5	500	100	500	1.5	20	70	50	70	30	50	15	100	150	50	200
	9051	3	1	.1	.2	500	70	500	2	15	50	30	70	30	30	15	100	150	30	200
Siltite	9042	3	.7	.5	.3	500	70	300	1	15	70	70	30	30	30	15	100	150	30	150

Silt-pebble	[6]9041	5	1.5	.7	.5	500	70	500	1.5	15	100	30	30	20	30	20	100	200	50	300
conglomerate.	[7]9043	5	1	.1	.5	500	100	500	1.5	15	70	30	50	15	500	15	N	200	30	150
Member E																				
Siltite	9050	1.5	.7	.1	.5	150	200	1500	1.5	5	70	L	50	15	20	15	100	150	50	300
	[8]9053	2	.5	.15	.3	300	70	500	1.5	20	50	5	50	20	20	15	100	150	30	200
	9055	1.5	.5	.2	.15	300	20	500	1.5	7	20	50	50	15	30	5	100	50	20	200
Quartzite	[9]9049	1.5	.7	.05	.3	150	150	700	1	15	50	L	30	15	10	15	N	150	30	200
	9052	2	.3	.15	.2	200	50	300	1.5	10	30	15	70	15	20	10	100	70	30	300
	9054	.7	.3	.05	.15	100	30	100	L	N	15	L	50	5	N	5	N	20	20	500
Member D																				
Argillite	[10]9059	3	3	.5	.7	700	1500	1500	3	15	100	50	N	50	50	20	100	200	50	300
	9060	3	1	.2	.3	500	100	700	3	7	70	L	70	20	50	15	150	150	30	200
	9061	3	1	.2	.3	700	100	700	3	7	70	5	50	30	70	15	L	100	30	150
	9062	3	1	.2	.5	700	50	500	2	7	70	L	70	30	30	15	L	100	30	300
	9063	2	.7	.15	.3	700	70	700	2	5	60	L	70	20	30	15	100	100	30	200
Member B																				
Argillite	9009	5	.7	.1	.3	300	70	700	1.5	15	70	50	50	20	50	20	100	150	30	200
	9010	5	.7	.1	.3	500	70	700	2	15	70	20	70	15	50	20	100	150	50	200
	9011	7	.5	.07	.3	500	100	700	1.5	20	100	70	30	30	50	20	100	150	50	200
	9012	7	.7	.15	.3	500	50	700	2	10	70	15	50	10	50	20	150	200	50	200
	9013	5	.7	.5	.3	500	70	500	2	15	70	30	50	20	70	20	150	200	50	200
	9014	5	.7	.1	.2	300	100	500	1	10	70	50	50	15	30	15	N	100	30	200
	[11]9015	5	.7	.1	.3	300	100	500	2	5	70	15	20	5	50	20	100	150	30	200
	9016	5	.7	.1	.3	500	100	700	2	15	70	20	70	20	50	15	L	150	50	200
	9017	3	.7	.1	.2	300	70	300	1.5	L	50	20	N	5	50	15	100	100	20	150
	9018	5	.7	.1	.3	300	150	700	2	5	70	L	N	20	30	15	100	150	30	300
Member A																				
Argillite	9064	3	1.5	1	.5	500	15	300	3	L	100	20	N	7	100	15	300	150	10	200
	9065	3	1.5	.3	.5	500	50	500	3	N	70	7	N	5	70	15	200	150	15	200
	9066	3	2	.7	.3	700	30	150	3	7	70	N	N	20	70	15	300	150	30	200
	[12]9067	3	1.5	.2	.5	500	1000	700	3	N	100	7	N	10	30	15	L	150	15	200
	[13]9068	2	.5	.1	.3	300	500	300	2	N	70	5	30	5	30	10	100	100	20	150

[1]Also contains less than 200 ppm Zn.
[2]Also contains 5 ppm Mo.
[3]Also contains 7 ppm Mo.
[4]Also contains 10 ppm Mo.
[5]Also contains 7 ppm Mo.
[6]Also contains less than 10 ppm Bi and less than 200 Zn.
[7]Also contains less than 10 ppm Bi and 3 ppm Ag.
[8]Also contains 1500 ppm As.
[9]Also contains 500 ppm As.
[10]Also contains 0.5 ppm Ag.
[11]Also contains less than 5 ppm Mo.
[12]Also contains 30 ppm Sn.
[13]Also contains less than 10 ppm Sn.

Geometric means by stratigraphic unit and by rock type are listed in table 6, along with geometric means by rock type from the quartzite facies near Pend Oreille Lake that are taken from Harrison and Grimes (1970). Inspection of these analyses shows some trends and some differences between members that reflect differences in their paleoenvironments.

Near Plains, Mg, Ti, Mn, B, Ba, Be, Co, Cr, Pb, Sc, and V are most abundant in argillite and least abundant in quartzite (table 6). These elements probably entered the depositional basin mainly combined in the lattices of degraded clay minerals (Hirst, 1962; McLennan and others, 1983, p. 214), though B and Be may have been adsorbed by the clays from water in the depositional basin (Harder, 1970; Herrman, 1970, p. H-1). The elements Cu, Cr, Pb, V, and Zn correlate with organic matter in many sedimentary rocks (Vine and Tourtelot, 1969), and some of these elements in the Prichard may also have been incorporated into the sediment in organic matter.

Argillite of member D differs from that of the other members in containing more B, Ba, Be, and Mn and less Cu and V (table 6). The low content of Cu and V probably reflects the paucity of carbonaceous matter in member D. The low content of both carbonaceous matter and iron sulfide, which accounts for the olive-gray color of the argillite of the member as compared with dark- to light-gray colors of argillite elsewhere in the Prichard, may have resulted from deposition in less saline or more oxygenated water. The higher concentrations of B, Ba, and Be in member D suggests that either the clays of the member were derived from a different source than the clays of the other members or that more of these elements were adsorbed by the clays from water in the depositional basin than was the case for the other members. More B, Ba, and Be could have been adsorbed if the clays were finer grained or if they were flocculated and deposited where fresh water carrying suspended clays encountered saline water (Herrmann, 1970, p. I-1, H-1). The relatively high Mn content of member D argillites suggests that the environment was more oxidizing than it was elsewhere during deposition of the Prichard Formation. In summary, minor-element abundance suggests that member D was deposited where river water encountered saline water that flocculated the clay.

GEOGRAPHY IN MIDDLE PROTEROZOIC TIME

All of the Prichard Formation is allochthonous (fig. 32), mostly as a result of Mesozoic thrusting, and data must be plotted on a palinspastic base before any conclusion can be made about the paleogeography. First, however, one must confront the proposal by Chamberlain and Lambert (1985) that the Purcell and Belt Supergroups were part of a microcontinent, Cordilleria, that in Early Cretaceous time lay in the latitude of California and was rotated 45°-50° counterclockwise of its present orientation. They consider the southern Rocky Mountain trench to be the location of the suture of Cordilleria with the North American continent. Price and Carmichael (1986) pointed out that there has been no right-lateral slip along the southern Rocky Mountain trench such as would be required by the proposal of Chamberlain and Lambert (1985). Furthermore, and fatal to the concept of Cordilleria, in the Little Belt Mountains of Montana, well east of the trench, the Belt Supergroup is in sedimentary contact with pre-Belt crystalline rocks (Weed, 1900, p. 281) and is thereby firmly riveted to the basement. Quite clearly, rocks of the Belt are native to the northwest. The Rocky Mountain trench in Montana, at least, rather than marking a mid-Cretaceous suture between terranes, is the result of normal faulting during Eocene and later extensions (J.E. Harrison, oral commun., 1986), though the position of the trench may reflect Precambrian events.

The Belt terrane has had a long and complex structural history that includes post-Belt but pre-Middle Cambrian folding and thrusting, Jurassic to Paleocene folding and thrusting, and Eocene and younger extension. Major strike-slip faults have formed and been reactivated during various structural episodes. Because of this complex history it is not yet possible to construct an accurate palinspastic map, but I have assumed that an approximate palinspastic reconstruction can be made by correcting the apparent offset on major strike-slip and thrust faults; Eocene and younger extension has been corrected only in part. I make no great claims of accuracy for the resulting map, but I am convinced that data plotted on the palinspastic base, inaccurate as the base may be, gives a much more realistic picture of the geography of Prichard time than would the same data plotted on the present geographic base.

The palinspastic base was made from the structural features shown on figure 32 by restoring the following fault displacements:

1. Twenty-six kilometers of right-lateral movement along the Hope fault near the Idaho-Montana border (Harrison and others, 1974, p. 12).
2. Twenty-six kilometers of right-lateral movement along the Osburn fault near Osburn (Hobbs and others, 1965, p. 77).
3. Six kilometers of horizontal movement across the Dobson Pass low-angle normal fault (Hobbs and others, 1965, p. 84).
4. Three kilometers of right-lateral movement along the Placer Creek fault (Hobbs and others, 1965, p. 81).

TABLE 6.—*Mean element content of rocks from the Prichard Formation*

[Geometric means (GM) and geometric deviations (GD) for samples from Plains area calculated from analyses in table 5; data from Pend Oreille Lake area from Harrison and Grimes (1970, table 7, p. O32–O33); numerical values substituted for N and L in table 5 are listed in Harrison and Grimes (1970, table 8); leaders (--), not determined]

Number of unit samples	Percent								Parts per million																													
	Fe		Mg		Ca		Ti		Mn		B		Ba		Be		Co		Cr		Cu		La		Ni		Pb		Sc		Sr		V		Y		Zr	
	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD	GM	GD
Plains area																																						
Argillite																																						
9 Member H	2.9	1.3	0.8	1.2	0.04	1.4	0.37	2.0	243	1.3	56	1.2	581	1.2	1.8	1.3	7.5	2.7	65	1.3	19	1.7	57	1.5	14	2.3	43	1.6	18	1.2	29	3.6	146	1.3	38	1.6	182	1.2
9 Member F	2.9	1.3	1.0	1.2	.26	3.5	.36	1.4	366	1.7	89	1.3	528	1.3	1.5	1.2	12.8	1.6	79	1.4	24	2.0	33	2.7	17	2.2	31	1.8	17	1.3	59	3.1	155	1.1	34	1.4	194	1.1
5 Member D	2.8	1.2	1.2	1.7	.23	1.6	.39	1.5	654	1.2	139	3.9	762	1.5	2.6	1.2	7.6	1.5	73	1.2	3.0	5.6	35	4.0	28	1.5	44	1.4	16	1.1	67	2.1	125	1.4	33	1.3	222	1.3
10 Member B	5.1	1.3	.68	1.1	.14	3.0	.28	1.2	387	1.3	84	1.4	581	1.3	1.7	1.3	9.7	1.9	70	1.2	20	3.3	26	3.3	14	1.8	47	1.3	18	1.2	76	2.3	147	1.3	37	1.4	202	1.2
5 Member A	2.8	1.7	1.3	1.7	.33	2.5	.41	1.3	482	1.4	102	6.2	342	1.8	2.8	1.2	1.8	2.4	80	1.2	3.5	7.7	4.8	2.8	8.1	1.8	54	1.7	14	1.2	140	2.4	138	1.2	17	1.5	200	1.0
[1]40 Total	3.3	1.4	.91	1.4	.13	2.9	.33	1.4	369	1.6	83	2.2	561	1.5	1.9	1.3	7.9	2.5	73	1.3	13	4.0	28	3.3	15	2.1	41	1.6	17	1.2	62	2.9	143	1.2	32	1.5	193	1.2
8 Siltite	2.0	1.5	.77	1.8	.24	2.2	.26	1.4	312	1.5	59	2.0	404	1.5	1.5	1.2	9.2	1.6	46	1.7	6.0	4.8	46	1.3	19	1.4	24	1.6	12	1.2	68	2.5	107	1.5	34	1.5	214	1.3
9 Quartzite	3.1	1.5	.32	1.4	.13	2.9	.21	1.4	192	1.5	43	1.7	309	1.7	1.1	1.4	5.6	2.2	31	1.7	5.8	2.9	44	1.4	11	1.5	19	2.6	8	1.2	66	2.9	69	1.8	24	1.2	265	1.4
Pend Oreille Lake area																																						
37 Argillite	2.7	1.3	.73	1.5	.087	2.2	.36	1.4	260	1.6	25	2.0	590	1.3	1.6	1.6	5.9	2.0	45	1.5	12	2.5	43	2.1	4.3	3.5	5.7	2.2	9.7	1.4	37	3.6	60	1.3	30	1.3	150	1.1
8 Siltite	2.0	1.5	.57	1.7	.090	2.1	.21	1.3	180	1.4	11	2.4	340	1.4	.81	1.8	4.7	2.9	26	1.5	10	3.3	34	1.7	6.4	4.8	6.6	2.1	5.9	1.2	--	--	41	1.3	21	1.3	140	1.2
8 Quartzite	.84	1.2	.18	1.5	.27	2.4	.18	1.2	190	1.7	7.1	1.4	210	1.7	.84	3.6	2.1	2.2	18	1.5	5.4	3.1	40	1.4	2.2	3.0	5.4	2.1	4.3	1.4	25	2.7	24	1.5	20	1.4	150	1.2

[1]Includes two samples from member G.

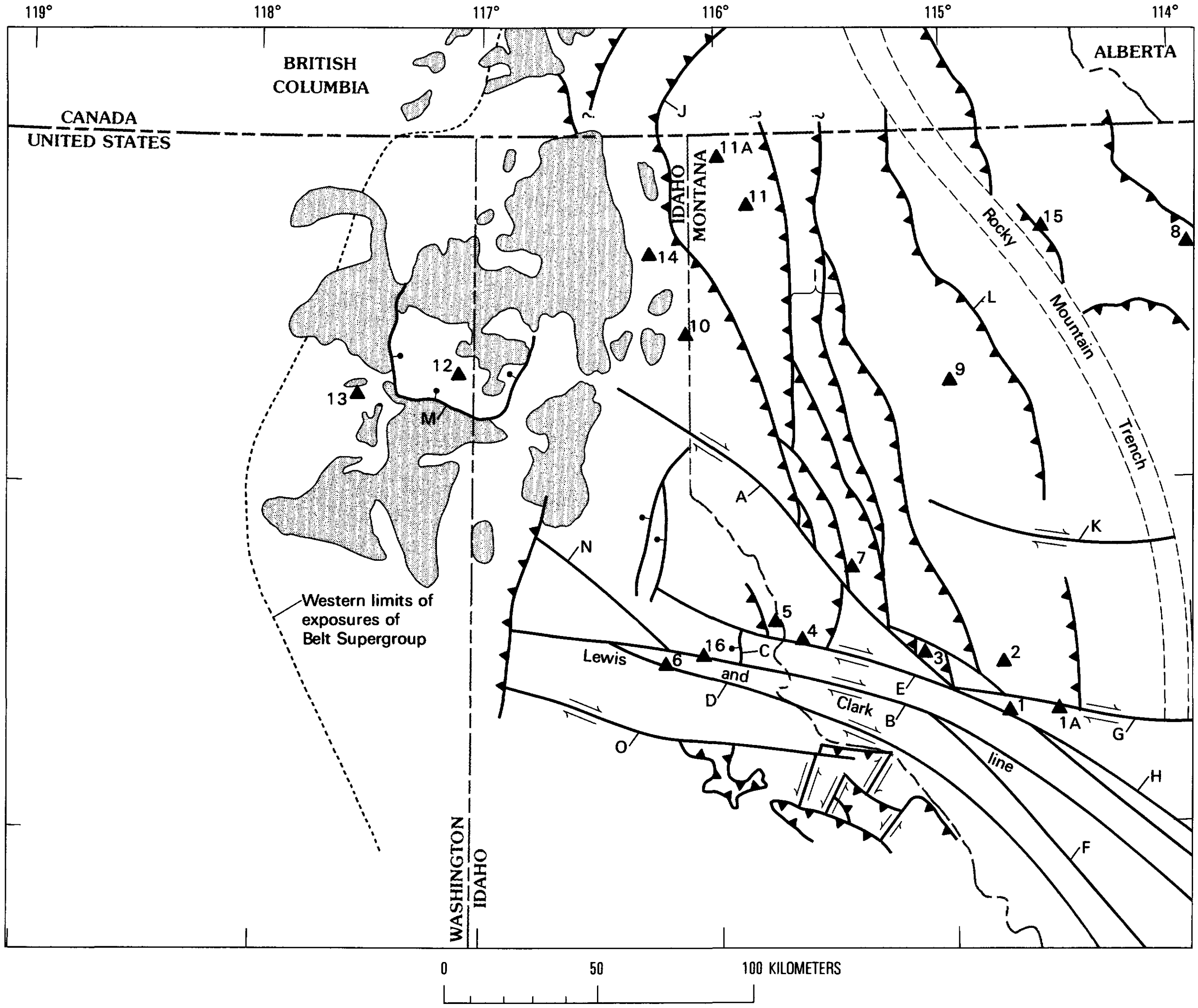


FIGURE 32.—Map showing major faults and intrusions in the western part of the Belt basin. Triangles are stratigraphic sections listed in table 1. Shaded areas are Cretaceous and Tertiary intrusives. Named faults are listed below. Data from Griggs (1973), Harrison and others (1986), Harrison and others (1983), Harrison and others (1972), Miller (1974a, b, c, d), Miller and Clark (1975), and Tipper and others (1981).

A, Hope fault
B, Osburn fault
C, Dobson Pass fault
D, Placer Creek fault
E, Thompson Pass fault
F, Boyd Mountain fault
G, St. Marys fault
H, Ninemile fault
I, Libby thrust belt
J, Moyie thrust
K, Big Draw fault
L, Pinkham thrust
M, Newport fault
N, Kellogg fault
O, St. Joe fault

5. More than 3 km of right-lateral movement along the Thompson Pass fault (Hobbs and others, 1965, p. 80).
6. Twenty-one kilometers of right-lateral movement along the Boyd Mountain fault (Hobbs and others, 1965, p. 81).
7. Six and one-half kilometers of right-lateral movement along the Osburn fault east of its intersection with the Boyd Mountain fault (Hobbs and others, 1965, p. 81).
8. Thirteen kilometers of right-lateral movement on the east end of the St. Marys fault (Harrison and others, 1974, p. 12).
9. Thirteen kilometers of right-lateral movement on the west end of the St. Marys fault (Cressman, 1985, p. 5).

10. Twenty-nine kilometers of right-lateral movement along the Ninemile fault (Wells, 1974).
11. Eleven kilometers of right-lateral movement along the Big Draw fault (from map relations in Harrison and others, 1986).
12. Shortening across the Libby thrust belt as measured from structure sections by J.E. Harrison (written commun., 1988).
13. The minimum shortening across the Moyie thrust belt.
14. Shortening across the Pinkham thrust and the thrust just east of the Rocky Mountain trench assumed to be the same as across individual thrusts in the Libby thrust belt.

In addition, it was assumed that the leading edge of the Lewis thrust sheet, nearly 50 km east of locality 8, figure 32, was derived from the position of the Rocky Mountain trench, as suggested by unpublished seismic sections, and that the Prichard Formation was transported 76 km eastward as shown by section A–A′ in Mudge and others (1982).

Relations west of the Moyie thrust are less certain than they are to the east. Immediately west of the Moyie, many normal faults have widened the outcrop belt of the Prichard Formation, but thrusts shown in Canada (fig. 32) may extend southward and shorten the section. The Newport fault is probably extensional (Harms and Price, 1983), but the amount of extension is not known. However, in the reconstruction, section 13 was moved slightly southeast with respect to

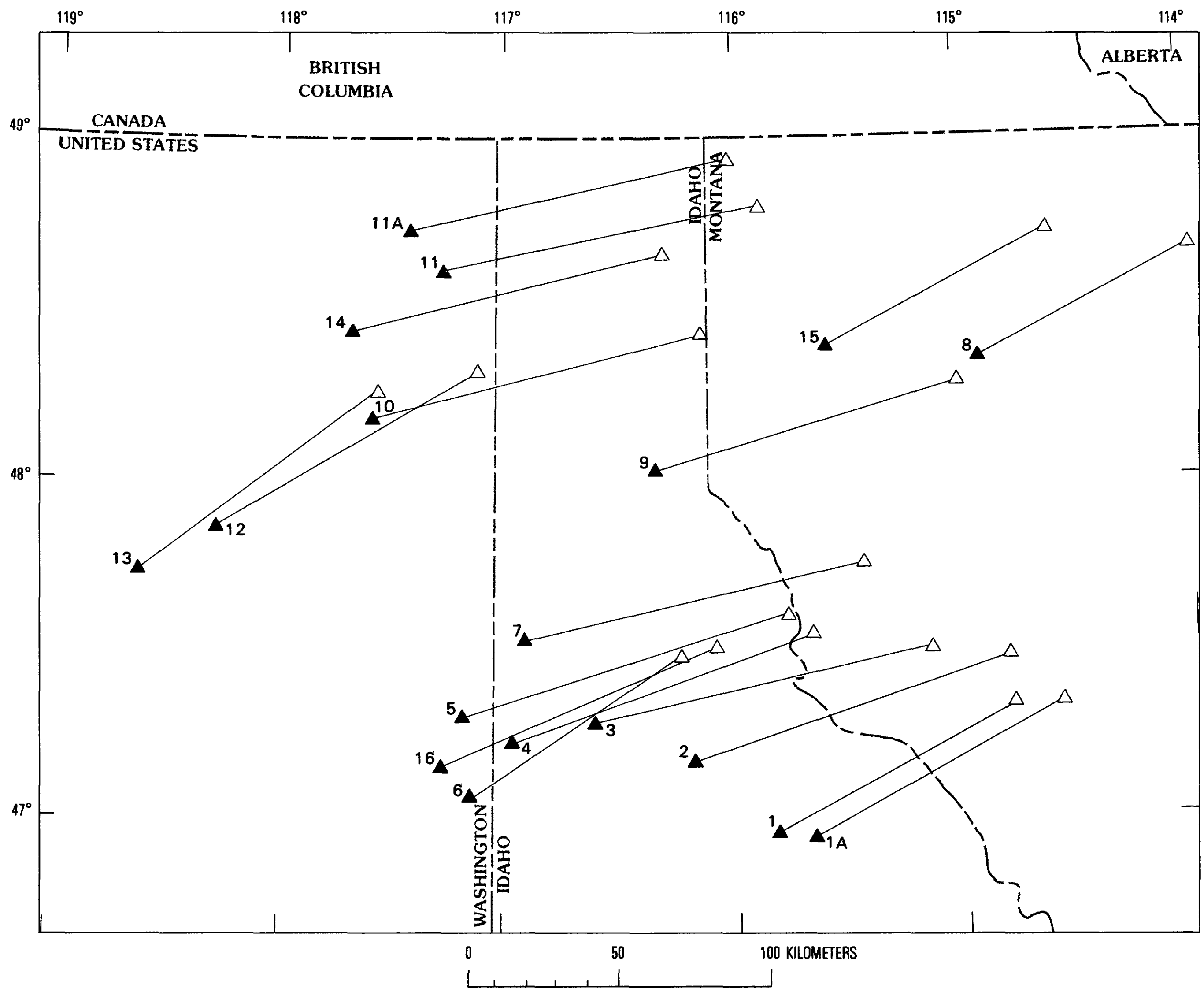


FIGURE 33.—Location of stratigraphic sections of Prichard Formation on approximate palinspastic base. Open triangles are present locations of sections; closed triangles are locations after palinspastic restoration.

section 12, and both were moved slightly eastward with respect to sections 10 and 14.

Palinspastic corrections for folding were not made within the fault blocks, and the positions of several fault blocks in the Plains area were adjusted to make more rational stratigraphic relations.

The positions of the stratigraphic sections before and after palinspastic reconstruction are shown in figure 33.

In addition to the translation that is restored in figure 33, some areas experienced considerable rotation. The differences in the orientation of fold axes north and south of the St. Marys fault suggest that those to the south were rotated about 40° counterclockwise (J.E. Harrison, oral commun., 1986) and plate 10 of Hobbs and others (1965) indicates considerable rotation, mostly counterclockwise, near the Osburn fault in the Coeur d'Alene mining district. Where directional features are plotted on the palinspastic maps, their directions have been rotated in these two areas to their inferred original orientation.

Several aspects of the geography during deposition of the Prichard Formation can be inferred from paleomagnetic data. Elston and McKee (1982, fig. 10) located the magnetic pole of 1,400 Ma at about lat 30° S., long 150° W., which would place the Belt basin near the equator, but Piper's (1982a, table 2) pole positions, which are based on more data, would place northwest Montana at lat 45° S. at 1,450 Ma and at lat 25° S. at 1,400 Ma. Thus, the Belt basin was probably in a temperate climatic zone during deposition of the Prichard Formation. Much of the time it may have been in the horse latitudes where rainfall is generally light.

A temperate climate in the source area is suggested by the chemical composition of fine-grained rocks in the Prichard Formation. The chemical index of alteration (CIA), defined as CIA= Al_2O_3 / $(Al_2O_3+CaO+Na_2O+K_2O)\times100$, for seven samples of argillite and siltite from the Prichard Formation, the analyses of which are in table 3, ranges from 67 to 78 and averages 72. Compared with the compilation by Nesbit and Young (1982), the indices of the Prichard samples are about the same as that of the average shale, higher than those of glacial sediments of both Early Proterozoic and Pleistocene age, and lower than those of muds of the Amazon cone. However, metamorphism may have changed the distribution of the alkali-metal and alkali-earth elements.

Finally, Piper's (1982a, table 2) poles place the Belt basin in Prichard time in the southern hemisphere. Thus, the Coriolis force, which was stronger in the mid-Proterozoic than today, would have deflected currents to the west.

LOWER SHALLOWING-UPWARD SEQUENCE

The argillaceous facies of the Prichard Formation comprises two shallowing-upward sequences, the lower consisting of members A through E and the upper of member F through the transition member. The depositional environments and paleogeography of the Prichard will be described in terms of these two sequences.

The distribution, thickness, and orientation of directional features of members A and B, C, and E are shown on the palinspastic base in figures 34, 35, and 36, respectively. Member D, for which no palinspastic map is presented, has about the same distribution as member E. The total extent of the members must be much greater than the known extent shown in figures 34–36. Member C is absent at Osburn, and member D, nearly 600 m thick near Plains, is probably less than 50 m thick at Osburn, but otherwise neither thickness nor facies trends in the exposed rocks suggest an approach to the margins of the units.

Members A and B, which total more than 1,600 m thick in their thickest exposure, consist mostly of graded siltite-argillite couplets that were probably deposited from low-density turbidity currents. Slump folds, common in member B, probably formed on the upper slope as a result of oversteepening by progradation (Cressman, 1985, p. 52–54). There is no direct evidence of water depth other than that it was below wave base, but the existence of at least 1,600 m of sediment devoid of shallow-water features suggests an initial depth of hundreds of meters. Slump directions indicate that progradation was to the southeast and that the source was somewhere westward. Data are insufficient to be more specific.

According to Drake (1976, p. 135–136), today less than 10 percent of the suspended load contributed to the sea by rivers reaches the deep sea; about 50 percent accumulates in estuaries and coastal wetlands and the remaining 40 percent is deposited off about two dozen major river mouths. However, the present is a time of high sea level and wide continental shelves. Much of the mud deposited on the shelves may be resuspended by the orbital motion of swells to form a turbid layer just above the sea floor, and if the shelf is narrow, as during a low stand of sea level, the material within the turbid layers may be transported across the shelf edge and deposited as widespread sheets of sediment on the upper slope (Sangree and others, 1978, p. 90). A similar mechanism is probably responsible for the silt and clay turbidites of members A and B. Nevertheless, the large amount of terrigenous material in members A and B suggests the proximity of the mouth of a large river.

The quartzite turbidites of member C seem to be part of a blanket deposit. The thinning upward sequences at the base of the member (fig. 8) suggest deposition in channels, but the member in its outcrop area near Plains extends at least 60 km in a direction nearly at right angles to the transport direction. The water was

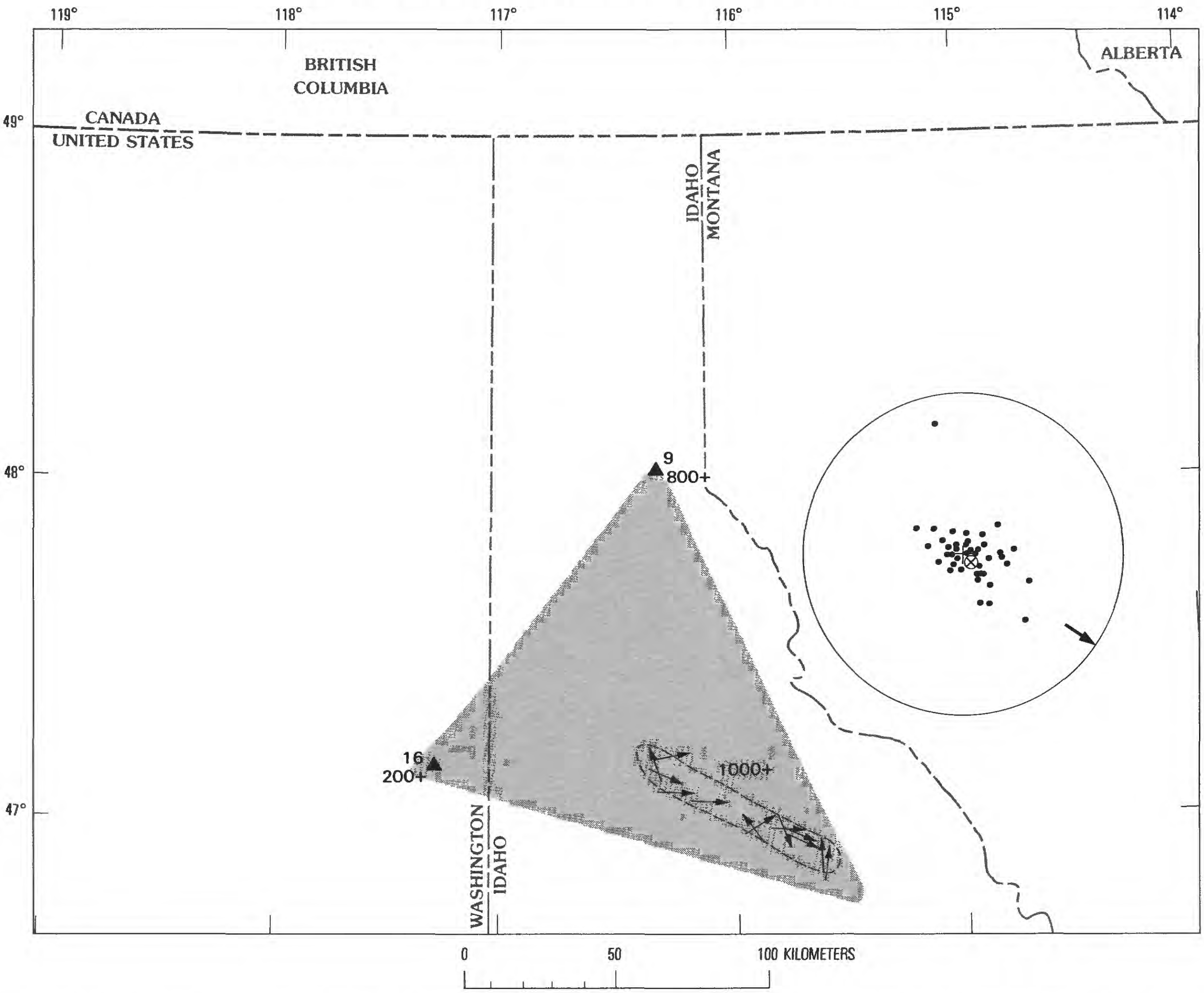


FIGURE 34.—Known distribution (shaded) and thickness of members A and B of Prichard Formation on palinspastic base. Dashed line encircles major outcrop area. Triangles are locations of sections 9 and 16. Maximum exposed thickness given in meters. Large circle is lower hemisphere stereographic projection showing poles to slump-fold axial planes, bedding rotated to horizontal; poles south of St. Marys fault (pl. 2) have been rotated 40° clockwise; x is mean pole; surrounded by 95 percent circle of confidence; arrow indicates mean direction of slumping. Small arrows indicate directions of slumping at individual localities.

sufficiently shallow, probably 200 m or less, so that long-period waves could form ripples on the surface of previously deposited turbidites. The sands were transported southwestward across the outcrop area. Inasmuch as the Belt basin was probably in the southern hemisphere and inasmuch as the Coriolis force, which deflects currents to the left in the southern hemisphere, was stronger then than now, the sands of member C were probably derived from the east.

The minor-element composition of the argillites of member D, discussed above, suggests that the member was deposited where river water carrying suspended sediment mixed with more saline water that flocculated the clays. Member D itself is a shallowing-upward sequence and is overlain by shallow-water sediments of member E. Member D is progradational, but the direction of progradation is not known. Evidence discussed below indicates that the lower part of member E was deposited at water depths of not more than 100 m and possibly not more than 50 m. These would have been the minimum depths for member D. The maximum water depth would have been the depth at which ripple marks formed in member C.

Member E is a shallowing-upward shelf deposit

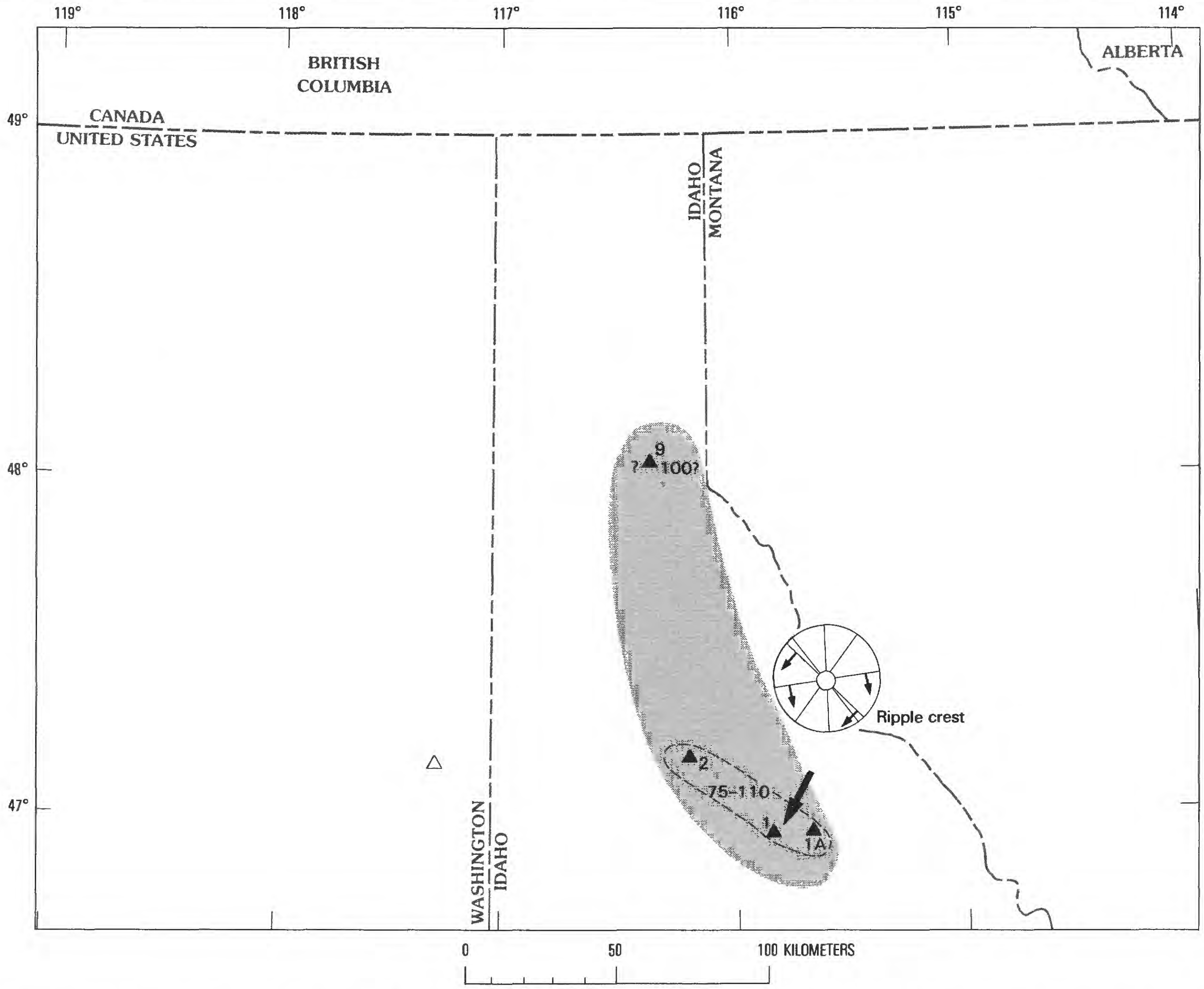


FIGURE 35.—Known and reasonably inferred distribution (shaded) and thickness of member C of Prichard Formation on palinspastic base. Dashed line encircles area of surface exposure. Solid triangles are locations of stratigraphic sections containing member C; query indicates probable presence; small numbers identify sections; open triangle is stratigraphic section where member D rests directly on member B. Large arrow is mean direction of sole marks (loc. A, pl. 2) rotated 40° clockwise. Ripple crests measured at locality B, plate 2; arrows indicate migration direction. Thicknesses in meters.

(Cressman, 1985, p. 51, 52). The maximum water depth is inferred from the presence of hummocky cross-stratification in the lower and middle parts of the section. At present, hummocky cross-stratification, which results from combined-flow storm waves, forms at depths of 100 m or less on open coasts and depths of 50 m or less where the fetch is limited (Swift and others, 1983; McCave, 1985). Inasmuch as land existed both to the east and west or southwest of the Belt basin in Montana during this part of Prichard time, the fetch was limited, and the maximum depth of 50 m applies. The minimum depth is indicated by desiccation cracks and mud-chip breccias in the upper part of member E near Plains that indicate subaerial exposure. The lenticular quartzite bodies nearly certainly originated as migrating sand ridges. Such ridges occur today at depths of 20–48 m on the New Jersey continental shelf (Stubblefield and others, 1984, p. 5); that seems an appropriate depth to postulate for these features in member E and is in accord with water depths indicated by the hummocky cross-stratification.

Although quartzites are conspicuous in member E, the unit consists overwhelmingly of siltite and argillite, and the sand bodies are encased in these fine-grained

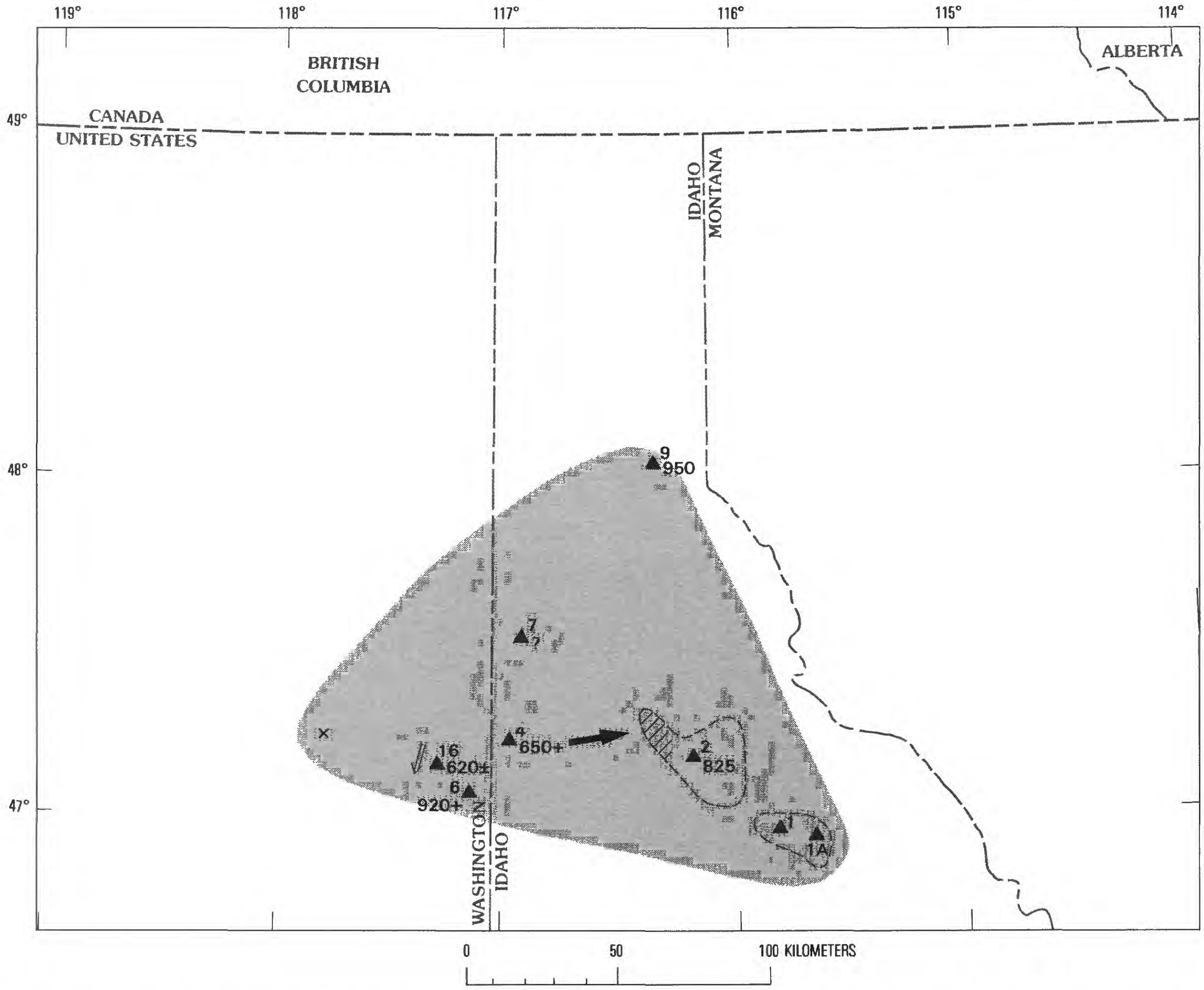


FIGURE 36.—Known distribution (shaded) and thickness of member E of Prichard Formation on palinspastic base. Dashed lines encircle major outcrop areas. Solid triangles show locations of stratigraphic sections that contain member E; small numbers identify sections; x is location of westernmost exposure. Lined area is inferred slump block; solid arrow shows direction of slumping. Hollow-shafted arrow is mean flow direction from flute marks near base of member rotated 30° clockwise. Thicknesses in meters.

rocks. Swift and Rice (1984) have proposed that similar sand bodies encased in shale in Cretaceous shelf deposits in the continental interior of the United States resulted from the concentration of minor sand in the mud by storm currents. However, the siltites and argillites in the Prichard contain very little sand-sized quartz, and a more likely explanation, proposed by Swift and others (1983) for sand ridges on the New Jersey continental shelf, is that shelf sand ridges form on a shoreface undergoing erosion. As the coast retreats, the sand ridge is separated from the beach and migrates down coast in response to along-coast currents.

Mud-dominated shelves require a combination of a large supply of mud and low fluid energy (Stanley and others, 1983). Inasmuch as most muds are contributed to shelves by major rivers, the source was most probably a major river. A large riverine source is also suggested by the absence of carbonates in the shallow-water sediments of member E. The surface water of the Middle Proterozoic ocean was saturated or supersaturated with calcium carbonate because of the high carbon dioxide content of the atmosphere, and the absence of carbonates in member E implies dilution by fresh-water drainage from land. The requirement of low fluid energy indicates that fair-weather waves were small.

In considering the lower progradational sequence as a whole, progradational sequences described in the Pleistocene of the Gulf Coast are of interest. Two types of progradational seismic facies have been recognized in Pleistocene sediments off the coasts of Texas and Louisiana (Sangree and others, 1978, p. 108–111). The first type, the sigmoid-progradational facies, consists of clayey muds, is especially common at the shelf margin, and is deposited during periods of rising sea level or subsiding land surface. More than 1,000 m of sediments may accumulate in this facies. Low-velocity turbid-layer currents probably play an important role in the deposition of these deposits. The second type, the oblique-progradational seismic facies, results from outbuilding from an upper, shallow-water to subaerial surface. The facies is interpreted to have formed during periods of relatively rapid supply of sediments and a standstill or a very slow rise of sea level.

By analogy with the Pleistocene of the Gulf Coast, members A and B of the Prichard Formation were deposited from low-density, low-velocity turbidity currents at the margin of a shelf that bordered a previously formed depression. As the sediment was deposited, the critical slope angle was exceeded, and slumps formed. The shelf prograded to the southeast, and the source was somewhere to the west. As subsidence slowed, shelf and near-shore sediments of member E, fronted by the flocculated muds that formed member D, prograded across the shelf. Although no direct evidence for the source of members D and E has been obtained, these units were nearly certainly derived from the same source as members A and B and they prograded in the same direction as the sediments of members A and B. The quartzite turbidites of member C were probably derived from the east side of the basin and were deposited by currents that may have flowed southward in a saddle between the southeastward prograding shelf and a shelf that prograded westward from the eastern margin of the basin, though the location and shape of the basin toward which the currents were flowing are unknown.

Member A is probably underlain by deposits of the lower slope and basin floor consisting of siltites and argillites of turbidite origin, argillites deposited from suspension, and perhaps some quartzite turbidites.

This interpretation is undoubtedly much simpler than the actual events. On modern shelves the distribution of facies has been profoundly influenced by changes in sea level, in particular, by those caused by glaciation and deglaciation. Similar changes of sea level must surely have taken place in the Middle Proterozoic, but their effects are not obvious, and most interpretations in this report assume changes in subsidence rates rather than changes in sea level.

UPPER SHALLOWING-UPWARD SEQUENCE

The upper shallowing-upward sequence consists of members F, G, and H and the transition member of the argillaceous facies and their lithic equivalents of the quartzite facies.

The contact between rocks of member E at the top of the lower shallowing-upward sequence, deposited at or near sea level, and the rocks of member F at the base of the upper shallowing-upward sequence, deposited well below wave base, is either sharp as in the Plains area and east of Kellogg or gradational through 1–2 m as in the Pine Creek area. This abrupt change from shallow to deep environments could have resulted from a sudden decrease in the supply of sediment during tectonic subsidence, a rapid rise in sea level, or a sudden increase in the subsidence rate. The upper sequence consists of thousands of meters of sediments that were deposited below wave base before shallow-water sediments are again encountered in the transition member. This great thickness of sediment suggests that there was no appreciable decrease in the sedimentation rate; it further suggests that the basin formed was hundreds and perhaps thousands of meters deep; otherwise, sediment accumulation and sea-level changes would have resulted in at least periodic sedimentation above wave base. Such depths would rule out sea-level rise as the cause of the change in environment, and an increase in the subsidence rate is nearly certainly the explanation.

The large inferred shelf-edge slump that involves mostly member E (fig. 36) (Cressman, 1985, p. 44–49) seems to have moved a little north of east. This suggests that the basin subsidence that initiated deposition of the upper sequences began along a north-northeast-trending axis somewhere east of the Plains area. Debris flows and contorted bedding present locally in the lower part of member F attest to relatively steep slopes at that time.

The thicknesses and directional features of member F and the dolomitic siltite member, member G and the quartzite member, and member H and the argillite and upper members are summarized on palinspastic base maps in figure 37, 38, and 39, respectively.

Member F is known to be present only in the southern part of the area (fig. 41), though it may also be present in the westernmost section near Chewelah, but thickness trends suggest that it extends beneath most of the rest of the area, and the lower member of the Aldridge Formation of British Columbia may be its equivalent. The dotted isopachs in figure 37 have been projected in a conservative manner, but they are conjectural.

The argillites and silty argillites of member F were deposited both from suspension and from low-density

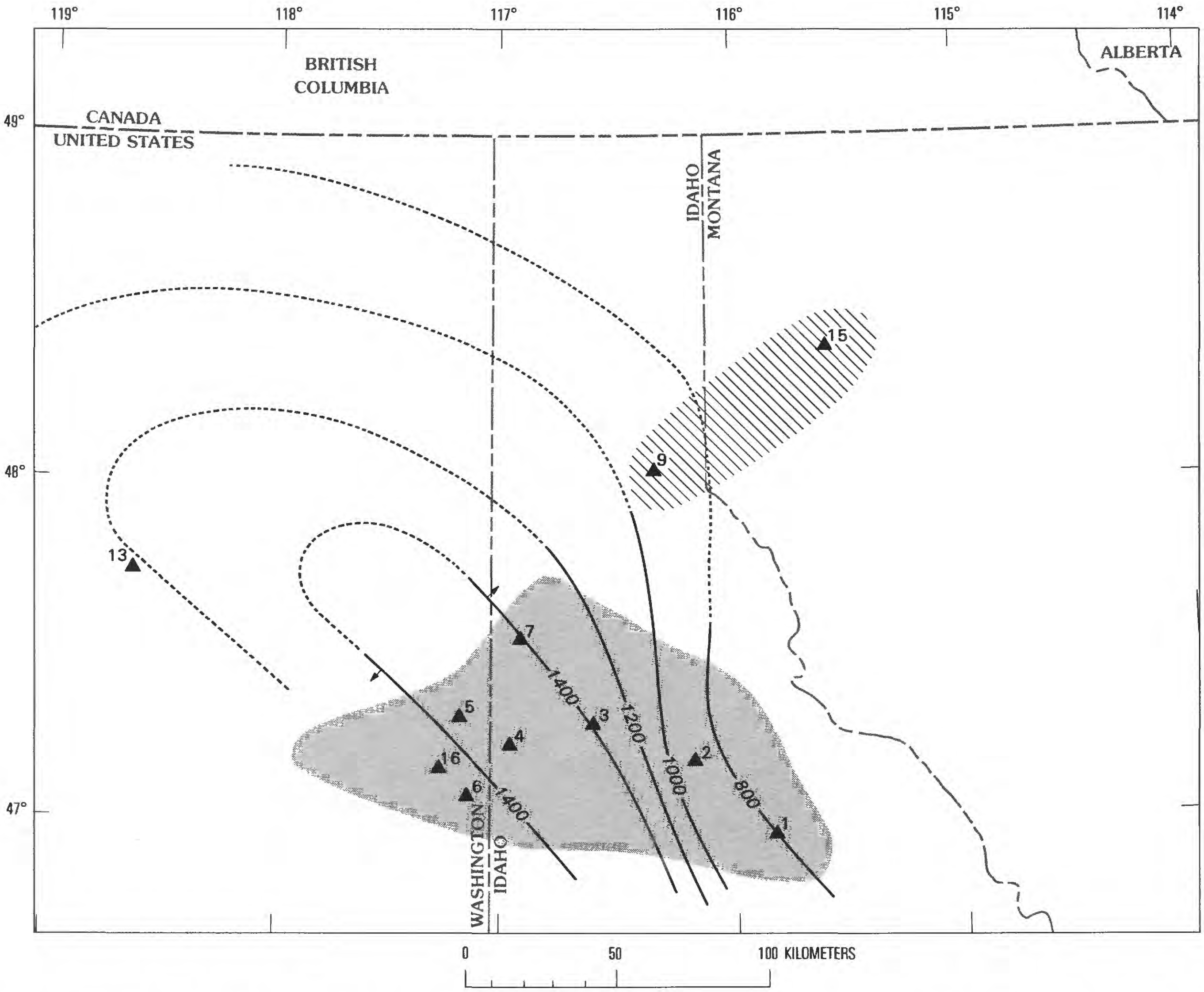


FIGURE 37.—Known distribution and thickness of member F (shaded) and dolomitic siltite member (lined) of Prichard Formation on palinspastic base. Triangles are locations of stratigraphic sections that contain the members; small numbers identify sections. Isopachs dotted where conjectural; isopach interval, 200 m.

turbidity currents. The stratigraphic position of the member below the quartzite turbidites of member G and at the base of the upper shallowing-upward sequence suggests that these beds were deposited on the sea floor in advance of the prograding submarine fan that comprises member G and the quartzite member. Member F contains abundant carbonaceous matter and abundant iron sulfides. In more recent sediments these constituents might suggest restricted circulation, but in Middle Proterozoic time when the oxygen content of the atmosphere was so much less and the oceanic mixed layer was thinner than today, carbonaceous matter could probably have accumulated anywhere below wave base if the organic productivity of the surface waters was sufficient, and sulfides could have formed by bacterial sulfate reduction nearly anywhere below wave base where the amounts of organic matter and dissolved sulfate were sufficient.

As discussed above, the member was deposited well below wave base. Actual depths cannot be determined, but the Sullivan submarine-exhalative lead-zinc deposit was probably formed under more than 400 m of water (Finlow-Bates and Large, 1978; Finlow-Bates, 1980, p. 156). This is based on an inference that the mineralizing fluid did not boil in the conduit and that the fluid was at a temperature of 250 °C and a salinity of 5 percent. However, in experiments by Bischoff and others (1981), sea water with a salinity of 3.7 percent did not

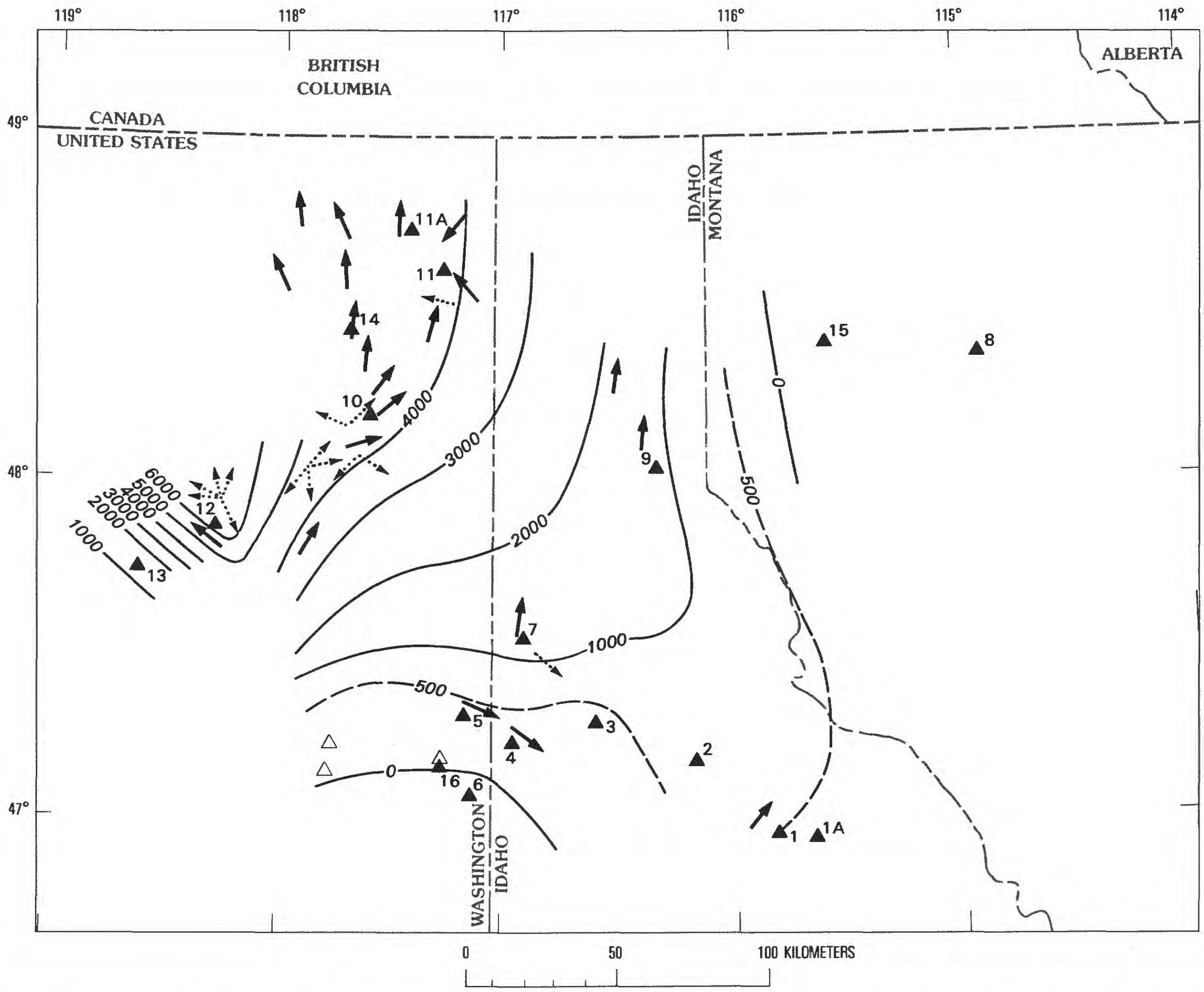


FIGURE 38.—Isopach map of member G and quartzite member of Prichard Formation on palinspastic base. Solid triangles are locations of stratigraphic sections containing the members; small numbers identify sections; open triangles are other localities where thicknesses were obtained. Solid arrows, average current directions from sole marks; some have been rotated as discussed in text. Dotted arrows show direction of slumping. Isopach interval, 1,000 m; 500-m isopach dashed.

remove heavy metal from graywacke, even at a temperature of 350 °C; on the other hand, brine containing 25 percent dissolved solids, though unable to complex heavy metals at 200 °C, removed significant amounts of the metals from the graywacke at 350 °C. If the mineralizing fluids at Sullivan were brines similar to those of the experiments and were at a temperature of 350 °C, the boiling-point curves of Haas (1971) indicate that the water should have been more than 1,300 m deep to have prevented boiling in the conduit.

The tourmalinized beds in member F required a source of boron, which most likely was clay in the surrounding sediment. Boron averages 130 ppm in clay and shale as compared with 35 ppm in graywacke and 20 ppm in limestone (Harder, 1970, p. K–1). Boron ranges from 230 to 2,500 ppm in saline clays (Harder, 1970, p. K–1), and any evaporite below the exposed section would have been a more than adequate source. Circulating pore water would have been required to scavenge the boron from the sediment and transport it to the site of deposition. Vigorous circulation is generally thermally driven, and the thermal gradient was therefore probably greater than normal. Inasmuch as the sediments were mostly clays and silty clays, circulation was probably along fractures, though none of these have been identified.

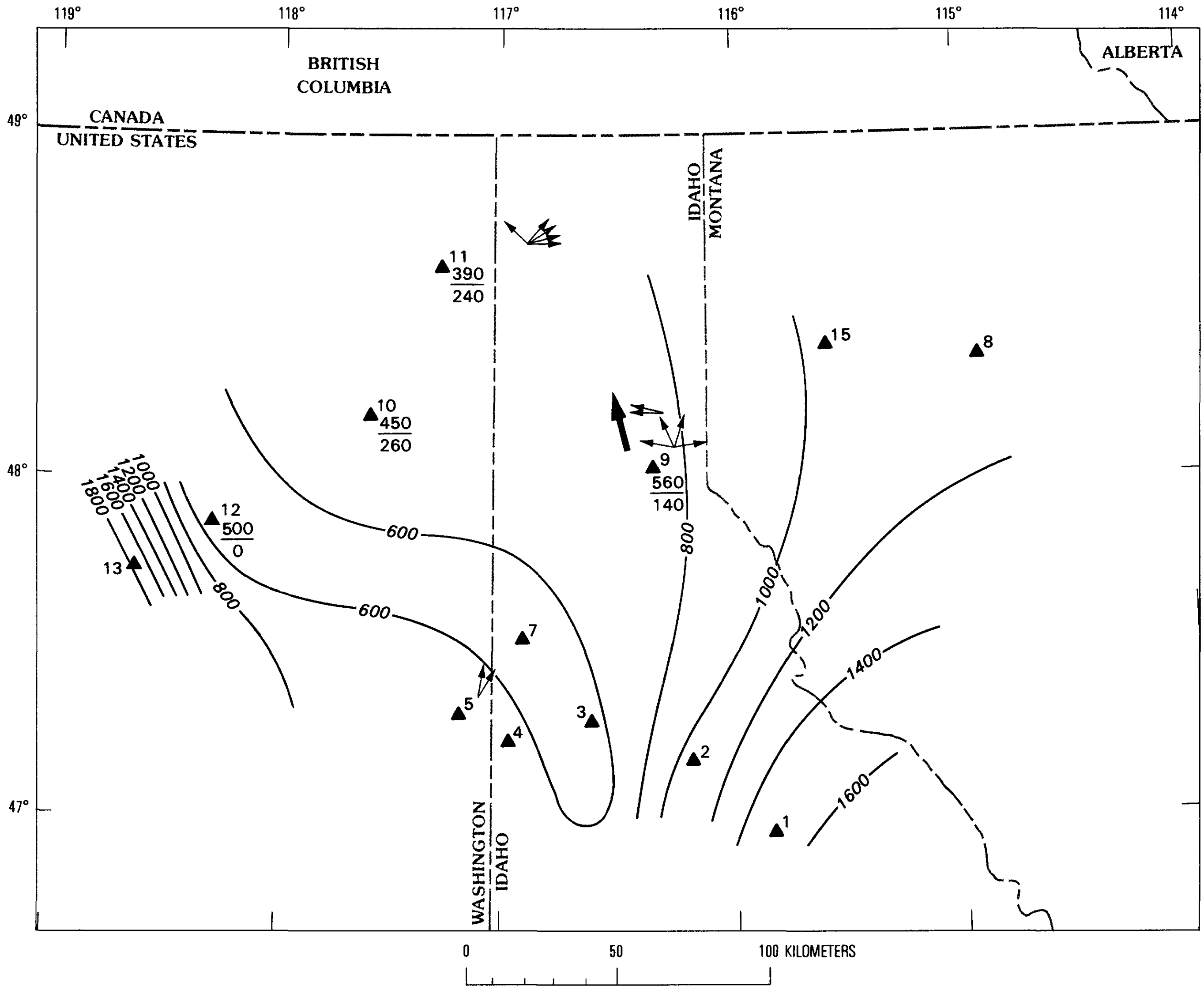


FIGURE 39.—Isopach map of member H and argillite and upper members combined of the Prichard Formation on palinspastic base. Triangles are locations of sections containing contoured units; small numbers identify sections; number below line gives thickness of argillite member; number above line gives thickness of upper member. Thin arrows show direction of slumping; heavy arrow is transport direction of quartzite channel fill. Isopach interval, 200 m.

In the Gibbs No. 1 borehole the dolomitic siltite member occupies the same stratigraphic position as does member H elsewhere. At Sage Creek in southern British Columbia just east of the Rocky Mountain trench and about 15 km north of the international border, greenish-gray dolomitic sandy argillite, which is probably an extension of the transition member of the Prichard Formation of this report, is underlain by 1,200 m of dolomite and limestone that contain only minor argillite (Fermor and Price, 1983, fig. 16). The carbonate section directly overlies the Lewis thrust. These carbonates fringed the eastern margin of the Belt basin in the latter part of Prichard time and were probably the source of the material in the dolomitic siltite member. The material was probably transported westward from the carbonate shelf by turbidity currents.

The quartzite member and its southeastern extension, member G, thin from more than 6,000 m in the west-central part of the area to a feather edge in the southern and eastern parts (fig. 38). North of lat 47°45′ N. and west of long 117° W. the base of the quartzite member is exposed only in the westernmost section, and the thicknesses shown are minimums. However, the bases of most of the sections are at about the same stratigraphic positions, and the trends of the isopachs are probably correct. I suspect that their positions are also nearly correct.

The current directions shown in figure 38 are averages for the directional features shown at each locality in plate 2. Directions have been rotated at locations south of the St. Marys fault near Plains and north of the Osburn fault in the Coeur d'Alene mining district as noted previously. The direction for the Bear Creek area (pl. 2) seems anomalous on figure 38; the measurements are from an area of complex structure immediately north of the Hope fault and may have been rotated, but I have no independent information on the direction or amount of any such rotation. Directional features north of Newport (pl. 2) define two directions, but all of the northeast directions are from one intensely faulted locality, whereas the northwest directions are from several localities. I have therefore taken the northwest direction as being correct.

Both the thickness trends and the current direction indicate that the quartzite was derived from a southern point source at about lat 47°15' N., long 118°15' W., and that the quartzite member was deposited as turbidites in a submarine fan, which is referred to in the remainder of this paper as the Prichard fan. The fanlike aspect is more apparent when data from the quartzite member's Canadian equivalent, the middle member of the Aldridge Formation, are added (fig. 40). The western part of the fan has not been preserved, but in figure 40 I have completed the contours so as to indicate the smallest fan allowed by the data.

A common approach in the study of ancient submarine fans has been that initiated by Mutti and Ricci-Lucchi (1972; translated from Italian to English by Nilsen, 1978) in which various facies are recognized and the positions of various facies assemblages on the fan are inferred. I have some difficulties in identifying the Mutti and Ricci-Lucchi facies, but in considering the sections on plate 3, most of the quartzite facies is most similar to the Mutti and Ricci-Lucchi facies C, the siltite-argillite intervals to facies D, and the slumped units to facies F; marker beds belong to facies G. Facies C and D are assigned to both the middle and lower fans in the Mutti and Ricci-Lucchi model, but in the absence of channel-fill quartzites all of the quartzite member and member G would be considered part of the outer fan. But are channel fills absent?

In studies of ancient turbidites coarsening-upward or thickening-upward sequences are commonly considered to have been deposited as prograding depositional lobes, whereas fining-upward or thinning-upward sequences are considered to have been deposited as channel fill. In the sections in plate 3 thickening-upward sequences mostly a few meters thick are common; thinning-upward sequences, also a few meters thick, are less so. This suggests that the member was deposited as thin, advancing depositional lobes and as shallow channel fills. However, channels detected by remote sensing on modern submarine fans are commonly 150–900 m deep and 18–150 km wide (Shanmugam, Damuth, and Miola, 1985, table 2), and features of this magnitude are too large to be detected in exposures of the Prichard Formation. Nevertheless, indirect evidence discussed below suggests that large channels and channel fills do exist.

Channels of modern deep-sea fans are commonly bordered by levees, and local slope failure is common, both on the inner slope of the levee into the channel and on the outer slope away from the channel (Nelson, 1984, p. 43). On figure 38 many of the slump directions are at a high angle to the direction of sediment transport. This suggests slumping of levee muds and therefore the presence of channels. Also, the two southwest transport directions shown on figure 44 suggest crevasse-splay deposits. Thus, some indirect evidence suggests the presence of large channels. However, neither large slumps nor anomalous transport directions are common.

The siltite and argillite intervals shown in the sections on plate 3 probably were deposited as overbank deposits on levees and in inter-distributary areas. These deposits closely resemble the silt-laminated mud described from midfan levee and overbank sites in the Mississippi fan (Dorrik and others, 1986). Some of the quartzite intervals must be channel fills, whereas others were probably depositional lobes, but I cannot distinguish one from the other. Channels on the midfan portions of modern deep-sea fans commonly range from 10 to 70 m thick (Barnes and Normark, 1985); presumably this includes the levees. Many of the quartzite intervals in the sections on plate 3 are within this range, especially if the effects of compaction are considered and if the overlying siltite-argillite intervals are included. Levees on the Indus fan are about 20 m high (Kolla and Coumes, 1985); many of the siltite-argillite intervals shown on plate 3 are of comparable thickness if the effects of consolidation are taken into account. The siltite-argillite units become thinner northeastward away from the point source, suggesting that levees and their channels became smaller in that direction.

The general picture that emerges from the above is of a large single submarine channel entering the area from the south and dividing into smaller channels that fanned out to the east, northeast, north, and northwest. Sand was deposited in channels, and silt and clay were deposited as overbank deposits; farther down the fan the channels became smaller, and much of the sand was deposited as depositional lobes. The channels may have migrated as in the Cap-Ferret fan (Cremer and others, 1985, p. 119), though in the southern hemisphere levee

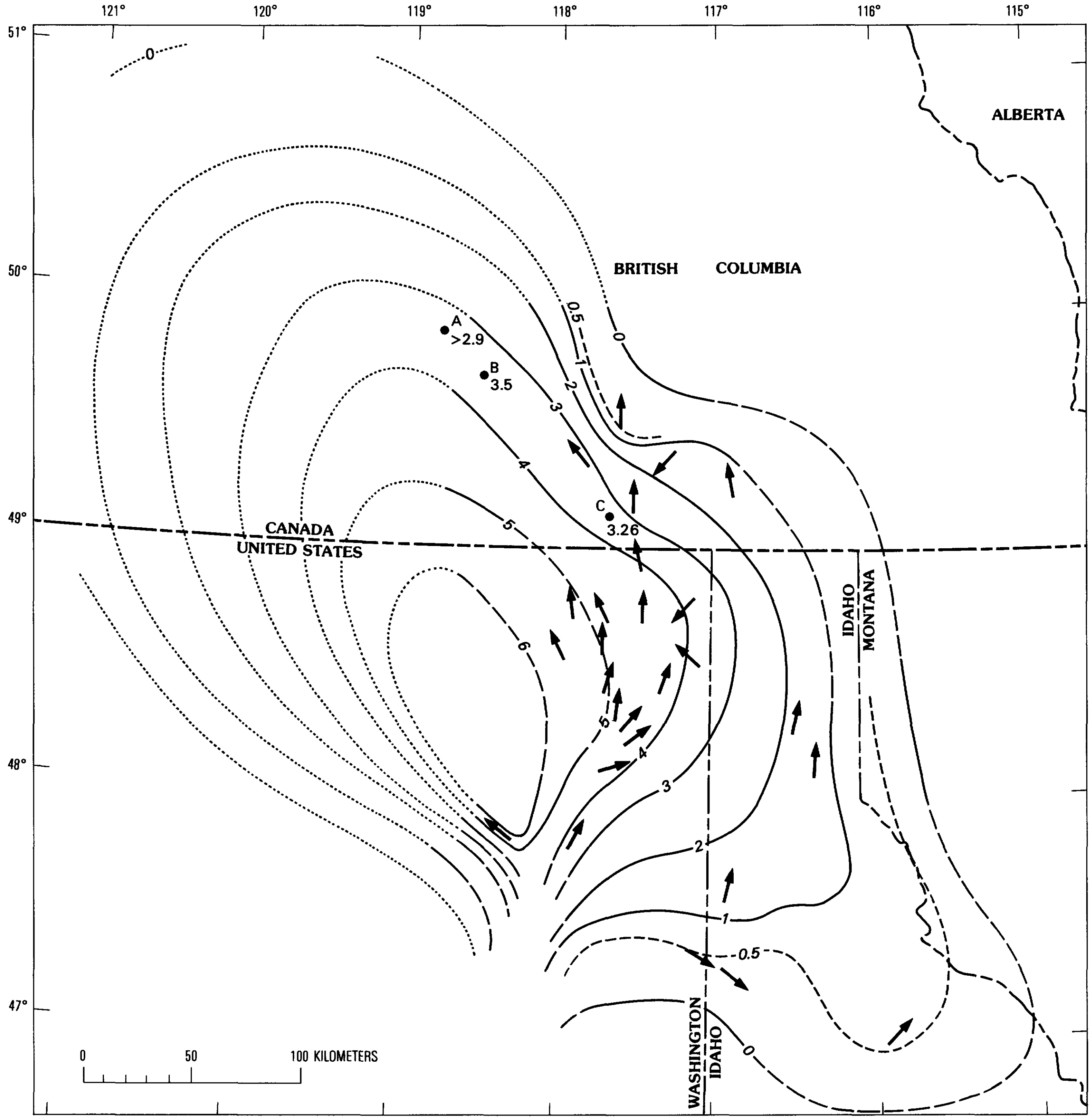


FIGURE 40.—Isopach map of member G and quartzite member of Prichard Formation in the United States and middle member of Aldridge Formation in Canada on palinspastic base. Data for United States from figure 38. Data for Canada: between long 117° and 118° W., transport directions from and isopachs modified from Høy (1984, fig. 3A); thickness at A from Reesor (1958), at B from Reesor (1973), and at C from Høy and Diakow (1982). Palinspastic reconstruction in Canada from structure sections by Price and Fermor (1984) and Okulitch (1984). Numbers give thickness in kilometers. Arrows are transport directions determined from sole marks. Isopach interval, 1 km; 0.5 km isopach shown locally. Isopachs long dashed where projected, dotted where conjectural. Conjectural isopachs drawn to include minimum reasonable area.

build-up would have been on the right-hand side and a channel would migrate to the left. The channels were probably periodically abandoned by avulsion, as on the Amazon fan (Damath and Flood, 1985, p. 104), and new channels were developed elsewhere.

A submarine fan is "a channel-(levee)-overbank system in which the overbank areas either consist primarily of muds or thin-bedded turbidite" (Bouma and others, 1985, p. 10). Therefore, not only is the western part of the Prichard fan missing, but so is the main channel and its flanking levees, which generally form the upper fan. The preserved part is probably equivalent to part of the middle and lower fans as identified in most of the modern fans whose characteristics have been summarized by Barnes and Normark (1985).

In figure 40 isopachs have been extended conservatively in order to compare the dimension of the Prichard fan with those of modern fans. The inferred size of the fan is nearly the minimum permitted by the data and, as noted above, it does not include the upper fan. The inferred area is 1.3×10^5 km^2, which is (1) larger than the areas of the present-day Monterey, Astoria, Rhone, and Magdelena fans; (2) smaller than the areas of the Mississippi, Laurentian, and Amazon fans; and (3) an order of magnitude smaller than the areas of the Indus and Bengal fans (Barnes and Normark, 1985). The volume of the Prichard fan calculated from figure 44 is 2.8×10^5 km^3 in its present state of compaction. Assuming the isopachs to represent the upper surface of the fan at the end of deposition and using the porosity-depth curve from Watts (1981, fig. 8), the volume at the end of deposition would have been 4.5×10^5 km^3. This volume, which excludes the upper fan, is an order of magnitude less than the volume of the Bengal fan, a little less than half that of the Indus fan, slightly less than the volume of the Amazon fan, and slightly more than that of the Mississippi fan (Barnes and Normark, 1985). Clearly the Prichard fan was a giant among fans, and it must have been deposited seaward of a very large river.

Presumably the rate of accumulation of the Prichard fan was comparable to that of modern large fans. From figure 31 it is estimated that the quartzite member was deposited through a period of 19 m.y. The quartzite member in the section shown in figure 31 is 4,110 m thick, which indicates an accumulation rate of 22 cm/10^3 yr for the present highly compacted state. Assuming an original bulk density of 1.5 g/cm^3, which is the average of eight samples from the Mississippi fan (Bryant and others, 1985, table 3), and 2.72 g/cm^3 for the present density of the Prichard, the rate of accumulation would have been 40 cm/10^3 yr. This compares with rates (1) on the Mississippi fan of 71 cm/10^3 yr for the Wisconsin interstadial, 6–11 cm/10^3 yr for the late Wisconsin glacial, and 2–13 cm/10^3 yr for the Holocene (Kohl and others, 1985, p. 272); (2) on the Amazon fan of 25 cm/10^3 yr (Damuth and Flood, 1985, p. 103); and (3) for mud during the Wisconsin on the Laurentian fan of 10–30 cm/10^3 yr (Piper and others, 1985, p. 140). This similarity of rates for present-day fans with that of the Prichard fan may be taken as evidence that the age span calculated for deposition of the quartzite member is reasonable rather than as an independent determination of the accumulation rate.

Changes in sea level have a profound effect on the growth rates of deep-sea fans. During high stands, such as that of the Holocene, most sediment brought to the ocean is deposited in estuaries or on the shelf, and at present less than 5 percent of the annual sediment input to the oceans reaches the deep sea (Collins, 1986). During low stands, when most of the continental shelf is exposed and rivers discharge at the shelf break, submarine fans experience rapid growth (Stow and others, 1985; Shanmugam, Miola, and Damuth, 1985). This is well illustrated by the present lobe of the Mississippi fan where rates of accumulation in centimeters per thousand years are as follows (Kohl and others, 1985, fig. 5):

	Middle fan	Outer fan
Holocene	7	2.5
Late Wisconsinan glacial	950	700

Converting to weight by assuming a porosity of 75 percent and a saturated bulk density of 1.47 g/cm^3 (Bryant and others, 1985, table 3) gives the following in g/cm^3/10^3 yr:

	Middle fan	Outer fan
Holocene	4.9	1.75
Late Wisconsinan glacial	665	490

The middle fan has an area of 4.5×10^4 km^2 and the outer fan has an area of 3.7×10^4 km^2 (measured from Bouma, Stelting, and Coleman, 1985, fig. 2), so during the late Wisconsin glaciation 4.9×10^{11} kg of sediment were deposited yearly on the middle and outer lobes, whereas during the Holocene a yearly average of only 2.8×10^9 kg have been deposited. Menard (1961) gave the volume of detritus contributed to the Gulf of Mexico by the Mississippi drainage since the end of the Jurassic as 11.1×10^6 km^3. Assuming an average porosity of 10 percent and a grain density of 2.8 g/cm^3, the total weight of this sediment is 2.8×10^{19} kg, and the average annual contribution of the Mississippi in the 144 m.y. since the Jurassic is 1.94×10^{11} kg. Thus, during the late Wisconsin glaciation all or nearly all of the load of the Mississippi was deposited on the fan, whereas at present a negligible amount reaches it.

Major Quaternary sea-level changes have been caused largely by glaciation and deglaciation, which seem to have resulted from orbital forcing (Bradley, 1985, p. 46). Variations in orbital characteristics presumably occurred in the Proterozoic as well, and glacial deposits are known from both the Early and Late Proterozoic. Furthermore, periodic sea-level changes have also been documented for the Mesozoic, a time for which there is no evidence of polar ice caps and abundant evidence of global climatic equability (Hallam, 1984, p. 235, 236). Therefore, by analogy with more recent fans the Prichard fan probably consists of turbidite lobes deposited during sea-level low stands separated by thin hemipelagic units, which are possibly represented by the banded argillite marker beds, deposited during high stands. As detailed measured sections and large-scale maps of the Prichard Formation become available, it should be possible to distinguish individual lobes, to determine their sequence in time and space, and to analyze the periodicity of changes in sea level.

The most common setting for deep sea fans today is on the continental rise in open ocean basins overlying either transitional or oceanic crust. Medium- to large-size fans are typically associated with rivers and are generally elongate (Nelson, 1984, p. 38, 40). Most of the large fans described in the volume edited by Bouma, Normark, and Barnes (1985) are under several thousand meters of water.

The shape of the Prichard fan was influenced both by the shape of the basin and by the Coriolis force that deflected currents to the left. The eastward margin of the turbidite basin was a west-facing submarine slope located in the position of the present Rocky Mountain trench. Turbidity currents entering the basin would have been deflected northward and northwestward by the slope, and this deflection would have been reinforced by the Coriolis force.

Member H, the argillite member, and the upper member, like member F, consist of silty argillites and argillites deposited from turbidity currents and argillites deposited from suspension, though the turbidites are probably the more abundant. The basal part of member H closely resembles member F, and like member F was deposited on the basin floor marginal to the Prichard fan. Slump folds common near the base of both the argillite member and the upper member suggest slumping of levee deposits, but most of the upper member, which is underlain by quartzite turbidites and overlain by shelf deposits of the transition member, must be a slope deposit. Sole marks in the one quartzite-filled channel exposed in the upper member indicate a current direction a little west of north (fig. 39), and the slope probably prograded in the same direction.

Taking 500 m as the thickness of the upper member and decompacting to the end of deposition of the member gives an original thickness of about 1,000 m. Assuming the shelf break to have been at a water depth of 200 m, the base of the slope would have been beneath 1,200 m of water if there had been no subsidence. Inasmuch as there certainly was subsidence, the base of the slope must have been considerably shallower, and the top of the fan as the upper member prograded across it was probably under a few hundred rather than a few thousand meters of water.

The massive silty argillite, slump-folded in part, that makes up the basal part of the transition member was a shelf-edge or uppermost slope deposit. The rest and by far the larger part of the transition member was a shelf deposit. The close interlamination of siltite and argillite, the lenticular and irregular laminae, the small-scale scour-and-fill structures, the cross-lamination, and the sparse ripple marks indicate varying current velocities and frequent reworking by waves. The parallel-laminated quartzites interbedded in parts of the member may be storm deposits similar to those described by Hayes (1967). The lighter color and the greater amount of quartzite in the upper part of the member indicates shoaling through time, though the unit did not aggrade to the intertidal zone except locally. The abundance of syneresis cracks suggests fluctuations in salinity that might be expected on a shelf in front of the mouth of a large river.

To summarize events during deposition of the upper shallowing-upward sequence, rapid subsidence formed a deep basin west of the Rocky Mountain trench, and muds were deposited on the basin floor from suspension and from low-density turbidity currents. Turbidity currents carrying a vast amount of sand, silt, and clay supplied by a large river to the south debouched on the basin floor and deposited a huge deep-sea fan that prograded northward and northwestward. Eventually the fan filled the basin to the point where slope and shelf sediments could prograde northward to nearly fill the depositional basin.

PRICHARD SEA

Geologists have speculated for nearly a century as to whether the Belt sea was part of the Proterozoic world ocean or whether it was an inland body of water (Winston, 1986, p. 114–119). There should be little doubt that during Prichard time at least the Belt sea connected with the world ocean. A river the size of the Mississippi draining into an enclosed basin and a fan the size of the Prichard submarine fan accumulating in a lake are not credible concepts.

SEDIMENT SOURCE

Holland (1984, p. 144–151) studied the effects of chemical weathering through geologic time by comparing the amounts of acid-titratable bases in sediments with their amounts in the parent rocks. The four argillites from the study area that were analyzed in table 3 contain 0.33–0.40 mol (MgO+CaO)/kg and 0.47–0.56 mol (FeO+ 2 Fe_2O_3)/kg corrected for size fractionation by Holland's method. These values, plotted on Holland's diagrams and compared with his analyses (Holland, 1984, table 5.6, figs. 5.6, 5.9), suggest that sediments of the Prichard Formation were derived by only moderate weathering from a terrane of granodiorite composition. That is, the source area consisted of normal upper-continental crust.

Dickinson and others (1983) plotted the composition of North American Phanerozoic sandstones derived from provenance terrane of various tectonic settings on QFL diagrams (Q, total quartzose grains; F, feldspar grains; L, unstable polycrystalline lithic fragments) and QmFLt diagrams (Qm, monocrystalline quartz grains; F, feldspar grains; Lt, total polycrystalline lithic fragments). The average composition of quartzite from the Prichard Formation in the Pend Oreille Lake area (table 2, column 1) and from the Aldridge Formation near Kimberly, British Columbia (table 2, column 2) were plotted on QFL and QmFLt diagrams and compared with the results of Dickinson and others (1983, fig. 1). If all the phyllosilicates in the Proterozoic quartzites were originally rock fragments, both analyses plot in the recycled orogenic field on the QFL diagram and in the mixed field on the QmFLt diagram. Contrariwise, if all the phyllosilicates are considered matrix, the analyses plot in the transitional-continental field on both diagrams. At least some of the matrix material is in the form of aggregates, and more must have been before deep burial and metamorphism; therefore, transitional-recycled seems a reasonable classification. That is, the source area would have consisted of both an older orogenic belt and part of the craton. Figure 41 shows that this was indeed the case:

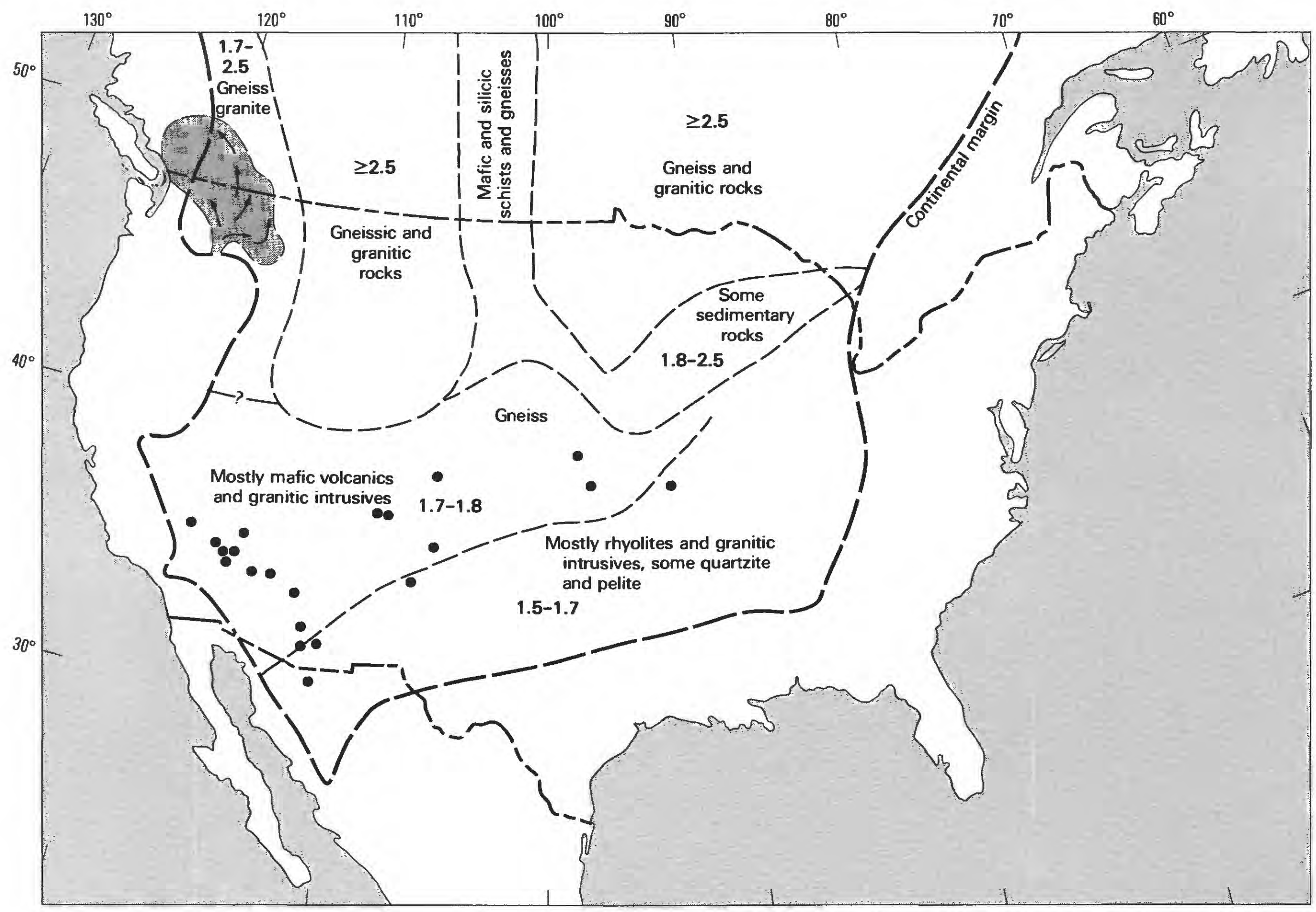


FIGURE 41.—Southern part of North American continent at 1.4 Ga, showing crustal provinces and crystallization ages in billions of years. Anorogenic granites intruded between 1.41 and 1.45 Ga are shown by dots. Shaded area is Prichard deep-sea fan from figure 40; arrows are transport directions. Data from Anderson (1983), Bickford and others (1986), Denison and others (1984), and Nelson and De Paolo (1985).

Archean gneisses to the south were bordered on their south by accreted volcanic arcs.

Bhatia and Taylor (1981, tables 5, 6) related the mineralogical and trace-element characteristics of mudrocks in the Tasman geosyncline of Australia to their plate-tectonic setting. The phyllosilicate-feldspar and phyllosilicate-quartz ratios for mudrocks of the Prichard Formation in table 2 are 1.7 and 2.7, respectively, for column 9 and 1.3 and 2.4, respectively, for column 11. The phyllosilicate-feldspar ratios of the Prichard mudrocks are most similar to the value for an active continental margin in the Tasman geosyncline, and the phyllosilicate-quartz ratios are intermediate between the values for active and passive continental margins in the Tasman geosyncline. Mudrocks of the Prichard (table 6) are most similar in La, La/Sc, and Sc/Ni to mudrocks from a continental-arc setting in the Tasman geosyncline, in Ni content to mudrocks from an oceanic-arc setting, and in Cr and B/Sr to mudrocks from an active continental margin. These comparisons can be explained if they apply to the source rather than to the depositional setting. The part of the North American continent south of the Belt basin consisted in Middle Proterozoic time of the south edge of the Archean craton and a series of island arcs that had been accreted to the craton in Early Proterozoic time (fig. 41; Condie, 1986).

The location of the sediment source is problematical. Frost and O'Nions (1984) studied the Sm-Nd systematics of some samples from the Belt Supergroup and concluded that rocks of the source terrane had separated from the mantle between 1.9 and 2.1 Ga. Inasmuch as rocks of the eastern and western parts of the Canadian shield give neodymium model ages of 2.6–3.2 Ga and 2.8–3.4 Ga, respectively, Frost and O'Nions concluded that the Belt sediments could not have been derived from the Canadian shield. However, their samples are small in number and from only a few localities, and they do not consider the effects of size sorting. Furthermore, a large area of the Precambrian in central United States has yielded crustal-formation ages of 1.7–2.0 Ga (Nelson and De Paolo, 1985, fig. 3). Even so, other considerations suggest that the North American craton could not have been the sole source, at least of the Prichard Formation.

Data for the lower shallowing-upward sequence are sufficient only to determine that the source for most of the material was somewhere to the west of the outcrop area, but data for the upper shallowing-upward sequence require a major river emptying into the Belt sea from the south. Figure 41 shows the location of the Prichard deep-sea fan with respect to that part of the southern North American craton that is older than 1.4 Ga. Given the location of the river mouth and the size of the craton at that time, it is most unlikely that the North American continent by itself could have provided a drainage area of the size required to supply sediment to the Prichard deep-sea fan. However, Sears and Price (1978) proposed that the Siberian craton was adjacent to the North American craton before 1.5 Ga, when rifting occurred. If rifting were as late as 1.4 Ga, the combined Siberian–North American craton might have been an adequate source. In fact, tectonic subsidence curves for the Paleozoic miogeocline of western Canada suggest that final separation did not take place until 555–600 Ma (Bond and Kominz, 1984).

Paleomagnetic data suggest not only that the Siberian and North American cratons were joined but that all the shields had aggregated in the Proterozoic to form a supercontinent (Piper, 1982a, b, 1983). Figure 42 illustrates that part of Piper's supercontinent that would have included the Belt basin. In this reconstruction a large drainage area would have been available to supply sediment to the Belt sea. The course and extent of the northward-flowing river is extremely speculative, but zones of weakness that subsequently were the loci of continental rifting are likely sites for major streams. As shown in figure 42 the Belt sea during Prichard time connected with the open ocean but was enclosed on three sides by land.

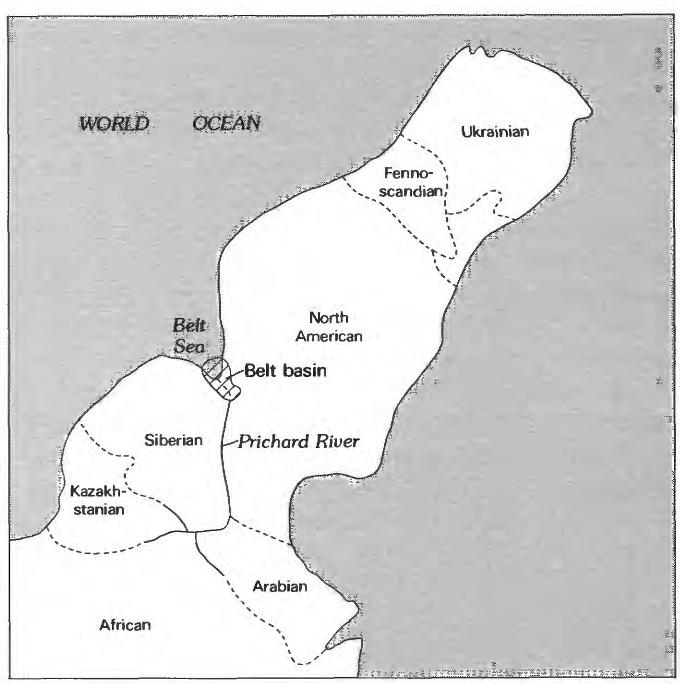


FIGURE 42. Part of Proterozoic supercontinent showing assemblage of shields according to Piper (1982b), location of Belt basin, and possible course of river that fed Prichard fan.

CHRONOCORRELATION

Finch and Baldwin (1984) proposed a chronocorrelation of the Prichard and Aldridge Formations based on the recognition of three major transgressive-regressive units that they termed lower, middle, and upper Prichard. Their publication is an expanded abstract, their stratigraphic columns are generalized, and locations are not given, and it is not easily apparent what they include in their units, especially in the quartzite facies. In the argillite facies the lower Prichard consists of members A through E, the middle Prichard of member F, and the upper Prichard of member G through the transition member. In the quartzite facies their middle Prichard seems to consist of the argillite member of this report and some of the underlying quartzite member; the lower Prichard is the lower half of the section and the upper Prichard all the beds from the top of the argillite member to the top of the section. Thus, as far as I can determine from their publication, they correlate the top of member E with a horizon several hundred meters below the base of the argillite member.

If the chronocorrelations of Finch and Baldwin (1984) are correct, much of the paleogeographic interpretation in this report is wrong, but I do not accept their correlations for the following reasons. First, I consider the Prichard Formation near Plains as having resulted from two major episodes of subsidence and basin fill rather than from three transgressive-regressive events as interpreted by Finch and Baldwin. Second, in the quartzite facies Finch and Baldwin interpreted some turbidites as being proximal and deposited in relatively shallow water and others as being distal and deposited in relatively deep water; the distribution of the "proximal" and "distal" turbidites was then taken as evidence for regression and transgression. However, in a submarine fan, turbidites with "proximal" characteristics are generally deposited in or at the mouths of submarine channels, whereas turbidites with "distal" characteristics are generally deposited in levees or in interchannel areas, and their distribution is not indicative of relative depth of water. Third, the paleogeography required by the chronocorrelation of Finch and Baldwin conflicts with that indicated by directional features. Their paleogeography requires a progradation in early Prichard time to the northeast. This is in obvious conflict with the data presented in figures 34–36 of this report.

It should be noted, though, that if I am correct, some of the sills in the argillite facies must have been intruded under as much as 2.5 km of cover.

Although I disagree with the chronocorrelation of Finch and Baldwin, I am unable to offer an alternative interpretation of my own. Key beds a through e certainly approximate time lines, and time lines certainly occur within the argillite member and the overlying tongue of the quartzite member, but I am unable to trace any of these throughout the basin. However, the analysis of the paleogeography in this report requires that the section below the top of member E be older than all of the quartzite facies.

PALEOTECTONICS

SUBSIDENCE IN PRICHARD TIME

The Prichard Formation resulted from two major episodes of basin subsidence and fill. The record of the early stages of the first episode is concealed beneath the lowest exposures, but the second episode is fully recorded by the upper shallowing-upward sequence. The second episode, and almost certainly the first episode as well, began with rapid subsidence that formed a deep basin that filled with sediment as the rate of subsidence decreased.

Figure 43 illustrates the amount of tectonic subsidence for that part of the upper shallowing-upward sequence of the Prichard Formation between the top of member E and the base of the transition member. The assumptions made in the construction of figure 43 were as follows: (1) an Airy-type crust; (2) present-day densities of 2.72 g/cm^3 for the sediment and 2.92 g/cm^3 for the sills (Harrison and others, 1972); (3) a density of 3.33 g/cm^3 for the Middle Proterozoic mantle; (4) the isopachs in figures 37 and 38 are correct; and (5) member E and the transition member were deposited at about the same depth of water. Figure 43 illustrates that tectonic subsidence for the interval considered was more than 3 km in the northeast and decreased both eastward and southward.

MECHANISMS OF TECTONIC SUBSIDENCE

Bott (1976) grouped explanations of tectonic subsidence under the headings of thermal-based and stress-based hypotheses. Thermal-based hypotheses include the following: (1) diapiric rise of hot asthenospheric material into the crust accompanied by uplift and followed by erosion, and finally, subsidence resulting from thermal contraction; (2) intrusion of the crust by substantial volumes of dense basic or ultrabasic intrusives; and (3) an increase of the density of the lower crust by metamorphism to the granulite or eclogite facies. The stress-based hypotheses include: (1) wedge subsidence near a mountain range as a result of imperfect isostatic adjustment; (2) thinning of the edge

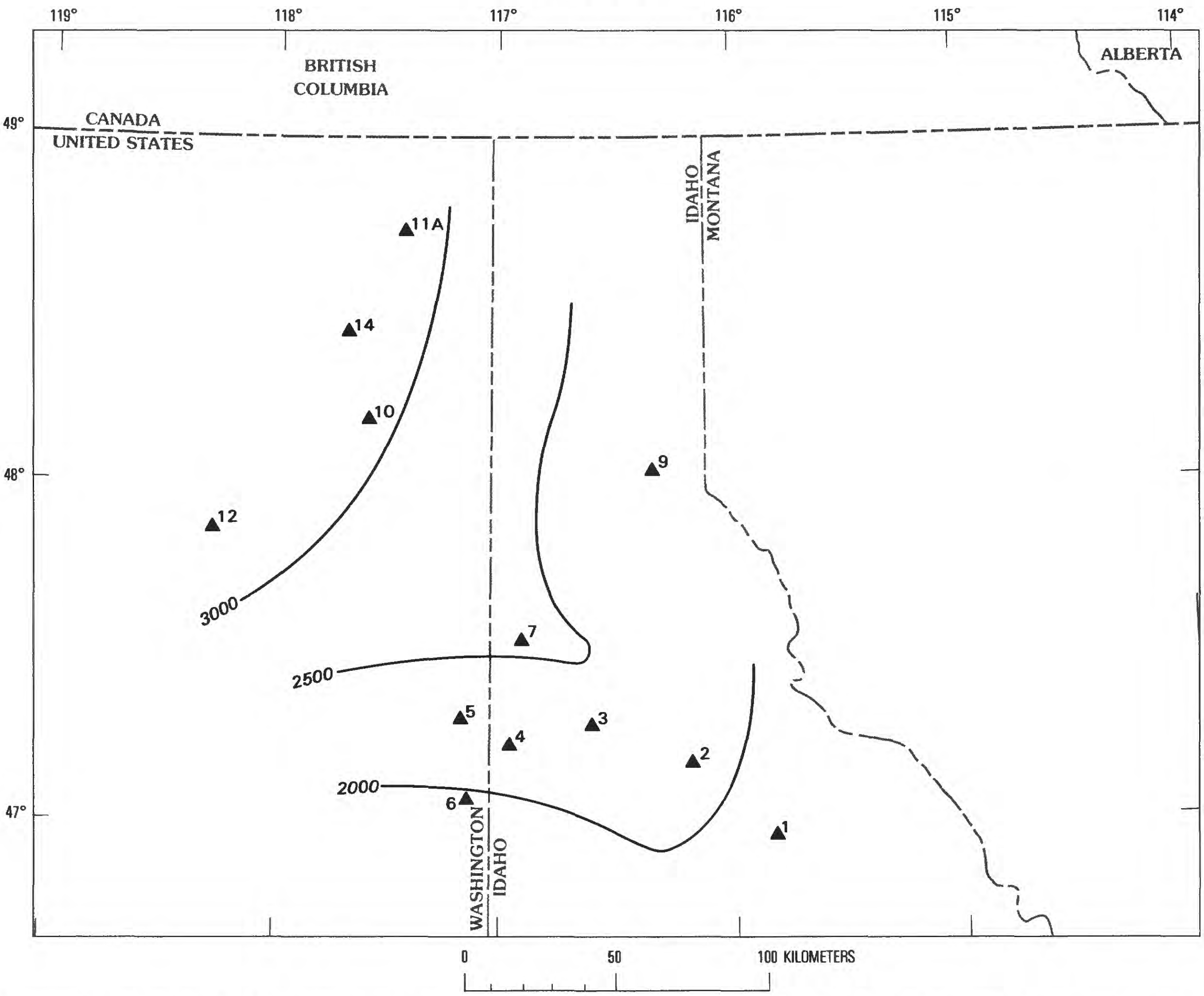


FIGURE 43.—Tectonic subsidence during deposition of upper shallowing-upward sequence of Prichard Formation. Contours in meters. Triangles show locations of section used in calculation; small numbers identify sections. Base is palinspastic.

of the continental crust and resultant subsidence of shelves at continental margins as a result of creeping of the crustal material toward the margin; (3) crustal stretching; and (4) local subsidence associated with transform faulting.

Continental rifting has commonly been explained by the thermal-based hypothesis of the rise of asthenospheric material, uplift, and subsequent subsidence during cooling. However, the Belt basin has neither the shape nor the dimensions of continental rifts. Furthermore, no evidence exists of either major boundary faults or of flanking uplifts, and the large amount of sedimentary fill, which must have been derived from a very large drainage area, suggests that an uplift that would have formed a drainage divide was not present on the southern basin margin. Additionally, alkali-rich magmatism characteristic of continental rifts (Bailey, 1983) is lacking and the sills that intrude the Prichard Formation consist of continental tholeiite.

Too little pre-Belt basement is exposed and the aeromagnetic data is too influenced by the magnetic properties of the Beltian section to know whether or not basic magma intruded the basement, but the gravity data suggest a granitic rather than dioritic composition (J.E. Harrison, written commun., 1987). I know of no way of evaluating the possibilities of phase change in the lower crust.

Of the stress-related hypotheses, the first, wedge subsidence near a rising mountain range, applies only to small basins. The inferred paleogeography for the Belt basin in Prichard time precludes shelf-edge creep as the primary explanation, and tensional basins in regimes of strike-slip faulting are generally much smaller than the Belt basin (see examples in Biddle and Christie-Blick, 1985). Thus, of all the explanations, crustal stretching in a tensional setting, possibly accompanied by intrusion by basic or ultrabasic magma, is the most likely cause of tectonic subsidence of the Belt basin.

Crustal stretching has commonly been a prelude to continental separation. Did stretching in Belt time proceed to actual separation, and was the Prichard Formation deposited in part on oceanic crust? Some evidence bearing on this question can be gained by examining the relation of the Prichard Formation to the crystalline basement.

RELATION OF THE PRICHARD FORMATION TO THE PRE-BELT BASEMENT

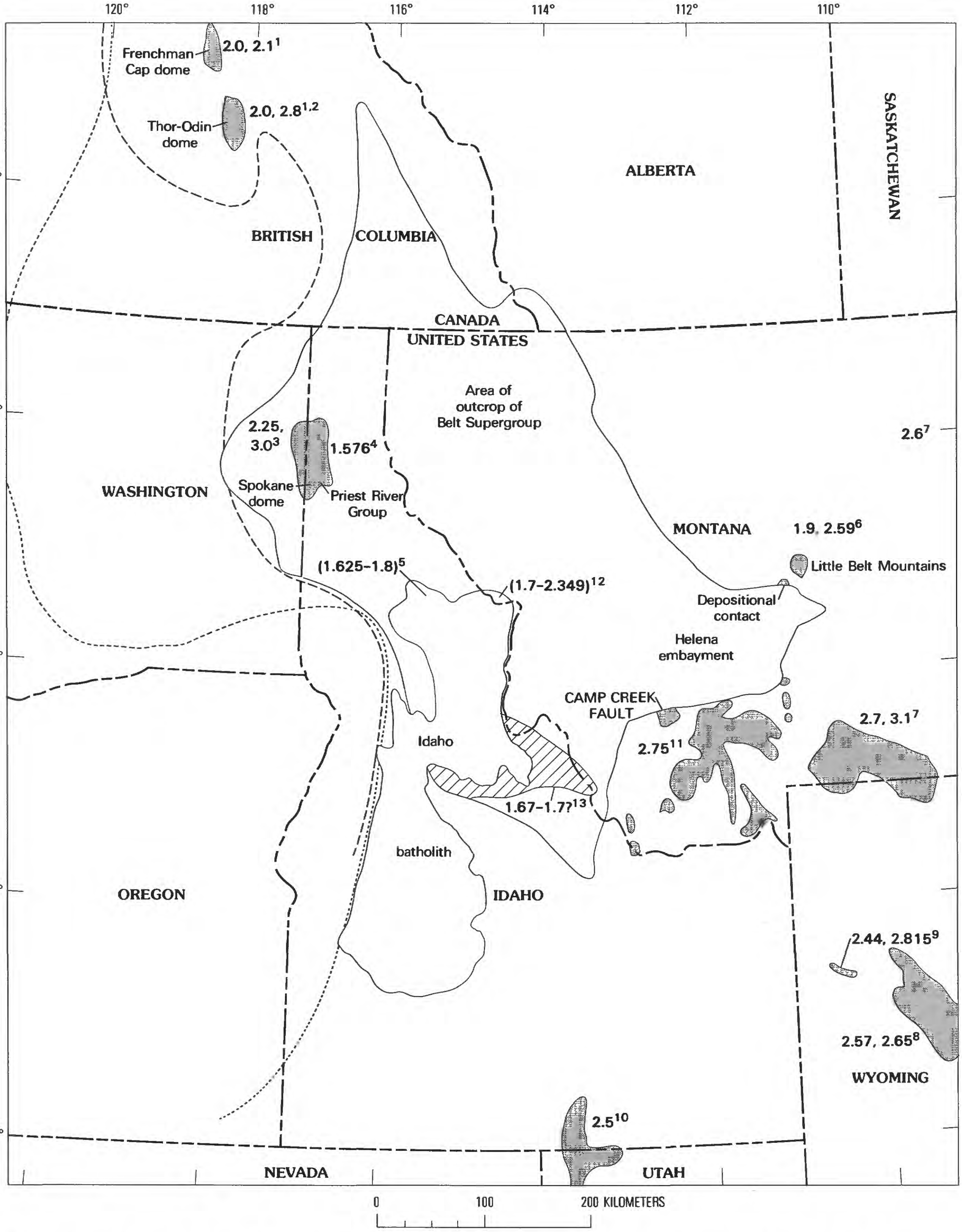
The entire Belt basin is underlain by continental crust (Armstrong and others, 1977), but the structural relations of the Belt to the underlying basement are known only very locally. In the Little Belt Mountains of Montana (fig. 44), the Neihart Quartzite, a probable temporal equivalent of some part of the Prichard Formation, rests in sedimentary contact on gneiss of Archean age, but elsewhere the contact between the Belt and the crystalline rocks is either faulted or concealed. On the south side of the Helena embayment the LaHood Formation is separated from Archean rocks to the south by faults (fig. 44), but the similarity in composition of clasts in the LaHood and the Archean to the south suggests that the LaHood was deposited in about its present position (J.M. O'Neill, oral commun., 1987).

In British Columbia, the Thor-Odin and Frenchman Cap domes and their overlying cover rocks constitute the Monashee Complex. The rocks of the Thor-Odin dome, Archean in age but intruded by Early Proterozoic orthogneiss, are overlain by a metamorphosed sequence of shelf-type sediments. At the base are 30–500 m of quartzite; these are overlain by 300–600 m of schist, gneiss, and lesser amounts of quartzite and marble; at the top are 300 m of calc-silicate gneiss and garnetiferous mica schist (Read, 1980). Reesor and Moore (1971, p. 112–113) tentatively suggested that the cover rocks correlate with the Lower Cambrian Hamill-Bagshot sequence, and Duncan (1984, p. 95) has dated these same rocks at 500–550 Ma by a $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ secondary isochron on syngenetic sulfides. On the other hand, Read (1980) correlated the cover sequence with the Belt Supergroup, but Brown and Read (1983) suggested that it is as old as 2 Ga. Both the Archean cores and the cover rocks have been folded and thrust in as many as six periods of deformation (Duncan, 1984, p. 123). The Monashee Complex is bounded by the Monashee decollement along which rocks of Late Proterozoic (Windermere Supergroup) to Jurassic age have been thrust eastward over the complex (Read and Brown, 1981). Obviously the true identity of the cover rocks has a considerable bearing on the paleogeography and history of the Belt basin.

The Priest River Group of western Idaho and eastern Washington, which is within the western part of the Belt basin (fig. 44), consists of mica schist, quartz-plagioclase schist, quartz-plagioclase-biotite gneiss, quartzite, and local pods of amphibolite (Griggs, 1973; Weiss, 1968). A zircon U–Pb age from the complex is 1.576±0.013 Ga (Evans and Fischer, 1986). Miller (1983) speculated that the Prichard Formation is in sedimentary contact with older gneiss of the complex on the east flank of the Selkirk Mountains where schists identified as Prichard are separated from the older rocks by coarse-grained pebble-bearing arkose. However, in my reconnaissance of that area I identified the schists as belonging to the upper part of the Prichard, and the contact with the older rocks cannot be depositional. Mesoscopic features in a mylonite carapace indicate that the Spokane dome, the southern part of the complex, has been overridden from west to east by the Prichard Formation (Rhodes, 1984). Thus, either the complex was part of the basement on which the Prichard was deposited or the Prichard was deposited west of the complex and thrust eastward. If the Prichard had been deposited entirely west of the complex, the easternmost exposures of the formation in Glacial National Park would have been transported eastward more than 250 km. Although not impossible, this is certainly unlikely.

Elsewhere the contact between the Belt Supergroup and the crystalline basement is concealed. Geophysical and structural data both suggest that in the area of this report crystalline basement is about 1,000 m below the lowest exposed beds of the Prichard Formation (J.E. Harrison, oral commun., 1988). The concealed basement from the Rocky Mountain trench east seems to be one block (Kleinkopf, 1984) and is undoubtedly part of the

FIGURE 44 (facing page).—Map showing relation of Belt basin to older rocks. Shaded areas are outcrops of pre-Belt basement; outcrop area of Yellowjacket formation is lined; numbers are crystallization ages in billions of years. Parenthetical numbers are ages of relict zircons from Idaho batholith of Mesozoic age. Dotted line is west edge of continental basement determined from $^{87}Sr/^{86}Sr$ ratio; line is at 0.704–0.706 transition (Armstrong, 1979; Armstrong and others, 1977). Dashed line is Late Cretaceous section between exotic terrane Quesnelia and North America (in Canada, from Okulitch, 1984; in United States, inferred from King, 1969). Sources of dates are as follows:



1. Parrish and Armstrong (1983).
2. Wanless and Reesor (1975).
3. R.L. Armstrong (oral commun., 1979).
4. Evans (1986).
5. Toth and others (in press).
6. Catanzaro and Kulp (1964).
7. Peterman (1979).
8. Naylor and others (1970).
9. Reed and Zartman (1973).
10. Compton and others (1977).
11. James and Hedge (1980).
12. Bickford and others (1981).
13. Hughes (1982), Hahn and Hughes (1984).

Wyoming Age Province of predominantly Archean age. On the other hand, the dates shown in figure 44 suggest that west of the trench the basement consists of rock of both Archean and Early Proterozoic age. Where these crop out, as in the Priest River Group and in the Frenchman Cap and Thor-Odin domes, there is no evidence of the presence of oceanic crust. Furthermore, the sills that intrude the Prichard Formation are continental tholeiites (Bishop, 1973) and presumably rose through continental crust. Thus, sparse evidence suggests that the Prichard was deposited on the continental crust that underlies it, but the crust must have been stretched to account for the existence of the basin.

OPENING OF THE BELT BASIN

Most mechanisms proposed to explain the development of basins by crustal stretching are variations on that first advanced by McKenzie (1978), which involves rapid stretching of the lithosphere and passive upwelling of hot asthenospheric material. The stretching causes block faulting of the upper brittle lithosphere, rapid subsidence to maintain isostatic balance, and slower subsidence as the upwelled asthenosphere cools.

McKenzie (1978, p. 29) related the amount of subsidence to the amount of stretching. Continental crust can be thinned to as little as 3 km (Montadert and others, 1979, p. 174). Assuming an original crustal thickness of 35 km, stretching to a thickness of 3 km would permit nearly 9 km of sediment to accumulate (McKenzie, 1978, fig. 4). This would account for either of the two basin-fill sequences that make up the Prichard Formation. However, if the entire Belt Supergroup resulted from one major episode of subsidence and basin fill, crustal stretching could not account for the entire decompacted thickness of the supergroup of nearly 19 km (fig. 3). This thickness would have required both crustal stretching and intrusion of the crust by basic magma as was inferred by Cochran (1981) for the northeastern Gulf of Aden.

Stretching results in normal faulting of the brittle upper crust, and syndepositional faulting might therefore be expected. Normal faults were active in the southern margin of the Helena embayment during deposition of the LaHood Formation (O'Neill and others, 1986). Høy (1978) and McMechan (1979) interpreted thickness and facies changes across two east-trending faults in the Rocky Mountains of southeastern British Columbia as indicating faulting in Belt time, though as noted by McMechan, the changes in stratigraphy across these two faults may instead have resulted from juxtaposition by postdepositional strike-slip faults. Aside from these occurrences, the ones in British Columbia being at least somewhat problematical, I know of no evidence of faulting concurrent with deposition of the Prichard Formation. If evidence for such faulting exists, it is in the concealed section between the lowest exposures of the Prichard and the basement.

The synsedimentary faults on the southern margin of the Helena embayment imply east-northeast and west-southwest stretching similar to the northeast-southwest opening proposed by Schmidt and Garihan (1986) on the basis both of the synsedimentary faults and of northwest-trending diabase dikes of Proterozoic age that were intruded into the Archean of the Highwood, Ruby, and Tobacco Root Mountains of southwest Montana. On the other hand, the faults described by Høy (1979) and McMechan (1979) would imply north-south extension.

Sills in the Prichard Formation are another indication of a tensional regime, and the orientation of the stress field might be inferred from the orientation of feeder dikes, but to my knowledge, no feeder dikes have been identified in the Prichard. However, the thickness and distribution of the sills may suggest the location and orientation of the feeders.

As summarized by Francis (1982), major dolerite sills of the world are nearly all tholeiitic in composition and are generally saucer-shaped if not deformed by later orogeny; the sills transgress section by step and stair, and dikes associated with the sills tend to follow a single preferred orientation in each region. Late Carboniferous tholeiitic sills in northern Britain are thickest in the bottom of sedimentary basins and are the result of (1) dike intrusion to a level where the density of the magma exceeded the density of the sediments, (2) lateral intrusion, and (3) downward transgression of the section to a level where hydrostatic equilibrium was established (Francis, 1982, p. 17, 18, fig. 26). The hydrostatic head then forced some of the magma up to produce the saucer shape. Thus, the immediate source of the magma was near the topographically highest part of the sill.

Sills in the Prichard Formation are tholeiitic, and where closely observable they cross section by step and stair (Cressman, 1981). They do not seem to have the saucer shape, but this may be because their total distribution is not known. Figure 45 shows contours on the top of two sills; datum is the base of the transition member, and the map is palinspastic. Although the contours are not tightly constrained and some of the correlations are uncertain, the crests of both sills are near the Idaho-Washington border and trend north-northwest. If Francis' (1982) hypothesis holds for these sills, the feeder dikes were along the crests and also trended north-northwest. Figure 46 shows the total thickness of sills in the Prichard Formation. If Francis' (1982) analysis is applicable, the feeder dikes were between the two maxima in figure 46 and flowed downsection to both the east and west. The data from the sills, then, suggest, but do not prove, that the sills were intruded along the Washington-Idaho border (on the palinspastic reconstruction), and figure 45 suggests that the feeder dikes trended north-northwest. If so, stretching was north-northeast and west-southwest.

The eastern edge of the stretched crust can be placed with some confidence at the Rocky Mountain trench. The Archean basement east of the trench is of normal continental thickness and produces subtle but distinctive northeast-trending gravity and magnetic anomalies (Kleinkopf, 1984) that could not have survived east-west attenuation and subsequent thickening by compression. The trend of the feeder dikes inferred above parallels the trench.

SUMMARY OF GEOLOGIC HISTORY

In Early Proterozoic time the southern margin of North America was a convergent margin, and from 1,800 to 1,650 Ma a succession of volcanic arcs and back-arc basins were accreted to the Archean nucleus of the continent (fig. 41) (Condie, 1986). However, in the Middle Proterozoic in the period between 1,500 and 1,400 Ma when the Prichard Formation was deposited the only activity was the intrusion of anorogenic granites into the Early Proterozoic terrane. If subduction continued, it left no clear record. In Montana a thermal event at 1,600±100 Ma that reset isotopic clocks in the Archean continental crust (Giletti, 1966) was about coeval with an episode of mantle enrichment in incompatible elements that probably resulted from the introduction of carbonate-rich melts (Dudas and others, 1987). This thermal event may have thinned the lithosphere by raising the subsolidus of the mantle material (Spohn and Schubert, 1982).

After 1.576 Ga, but before 1.433 Ga, divergence between the present North American craton and a large area of continental crust to the west, presumably the Siberian shield, resulted in stretching of the continental lithosphere, which was possibly already thinned by the preceding thermal event, in an east-northeast and west-southwest direction. The stretching occurred mostly west of the Rocky Mountain trench. Stretching caused faulting of the brittle upper crust, and the thinned crust was intruded by basic magma. Isostatic adjustment of the thinned and intruded crust resulted in subsidence that formed a gulf opening north-northwestward (in terms of present compass directions) to the world ocean.

The first sediments deposited on the newly subsided crust are concealed, but they may be a transgressive sand similar to the Neihart Quartzite or they may be conglomerate similar to those of the LaHood Formation and shed from basement fault blocks. Evaporites have commonly been deposited during early stages of continental separation, and some may have formed very early in the history of the Belt basin.

Isostatic adjustment to crustal distension is generally rapid, and in the Belt this initial subsidence formed a deep-water basin. Members A through E of the Prichard Formation are a shallowing-upward basin-fill sequence. The section below the lowest exposed beds of member A probably consists of fine-grained turbidites deposited at depths of hundreds to perhaps thousands of meters. In the Gulf of California, turbidites have been found resting on conglomerate that rests on granite (Saunders and Fornari, 1982); the unexposed basal Belt may be similar.

After isostatic equilibrium had been attained, slower tectonic subsidence resulted from cooling and contraction of upwelled asthenosphere. This tectonic subsidence decreased as a function of the square root of time. As the tectonic subsidence decreased, the basin was filled by fine-grained sediments that prograded mostly from the west. The absence of carbonates in even the shallow-water sediments suggests that the material was brought to the basin by a large river, but the location of the river and its mouth cannot be determined. Deposition of the shallow subtidal to intertidal sediments of member E closed the first episode of basin filling. Data on this first sequence are too sparse to determine the size and configuration of the basin.

Following the first episode of basin filling, the two shields again diverged to produce a second episode of crustal stretching, and the area again subsided to well below wave base. The crustal stretching and subsidence locally resulted in large-scale slumping of sediments deposited in the first episode of subsidence and fill. The newly subsided gulf was bordered near the Rocky Mountain trench by a shelf on which carbonate rocks were deposited. A major river emptied into the gulf from the south to form a huge deep-sea fan. A sketch map showing the inferred paleogeography at that time is presented as figure 47. Many parts of the map are conjectural, including the location of the eastern and southern margins of the sea, but it illustrates the concepts developed in this paper. As the rate of tectonic subsidence slowed, the basin was again filled, and slope and shelf sediments prograded northward across the area to terminate deposition of the Prichard Formation. Deposition of mostly shallow-water to subaerial sediments continued throughout Belt time.

Following deposition of the Belt Supergroup the basin was subjected to east to northeast compression during the East Kootenay orogeny as redefined by McMechan and Price (1982). If the inferred paleogeography is

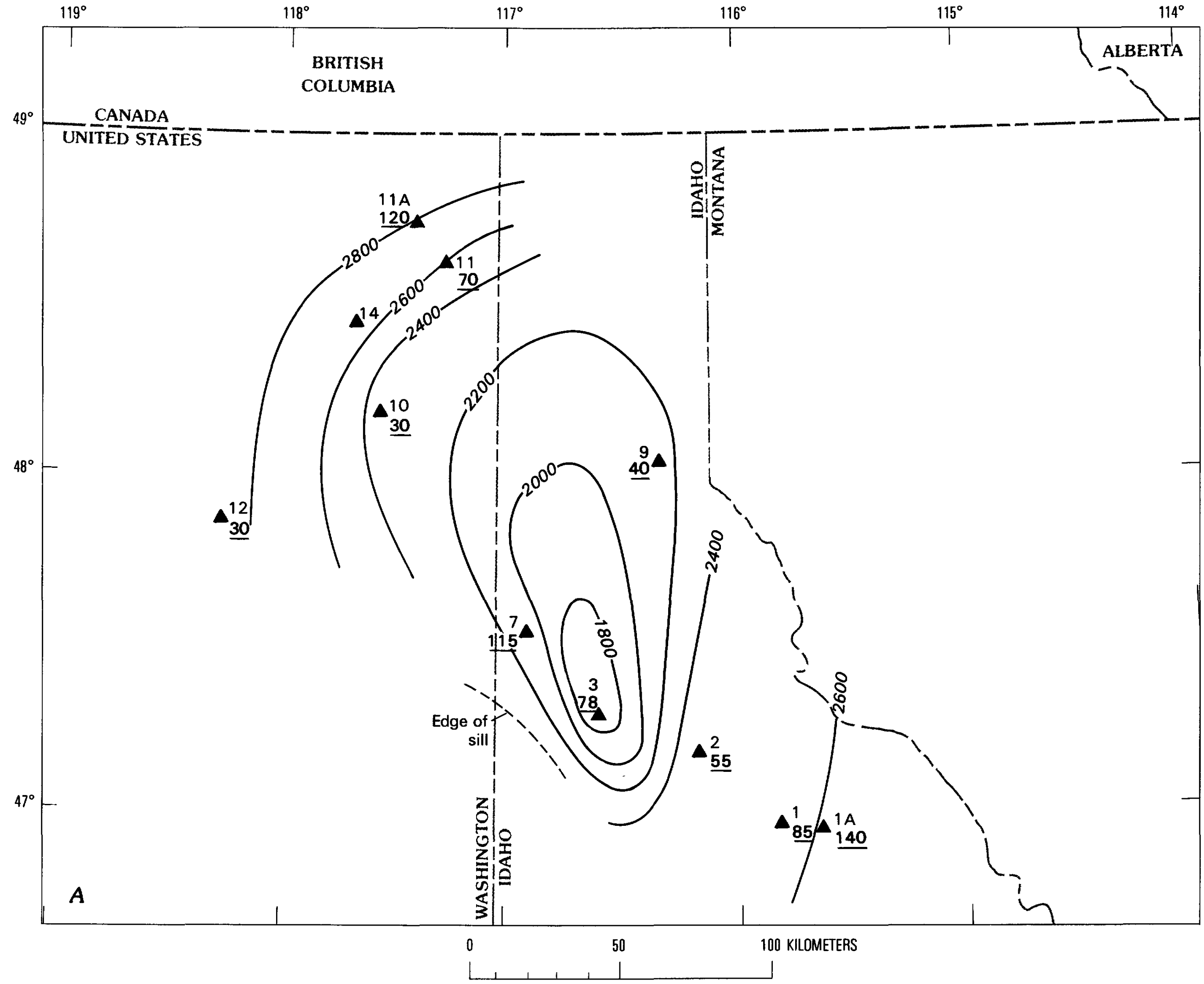
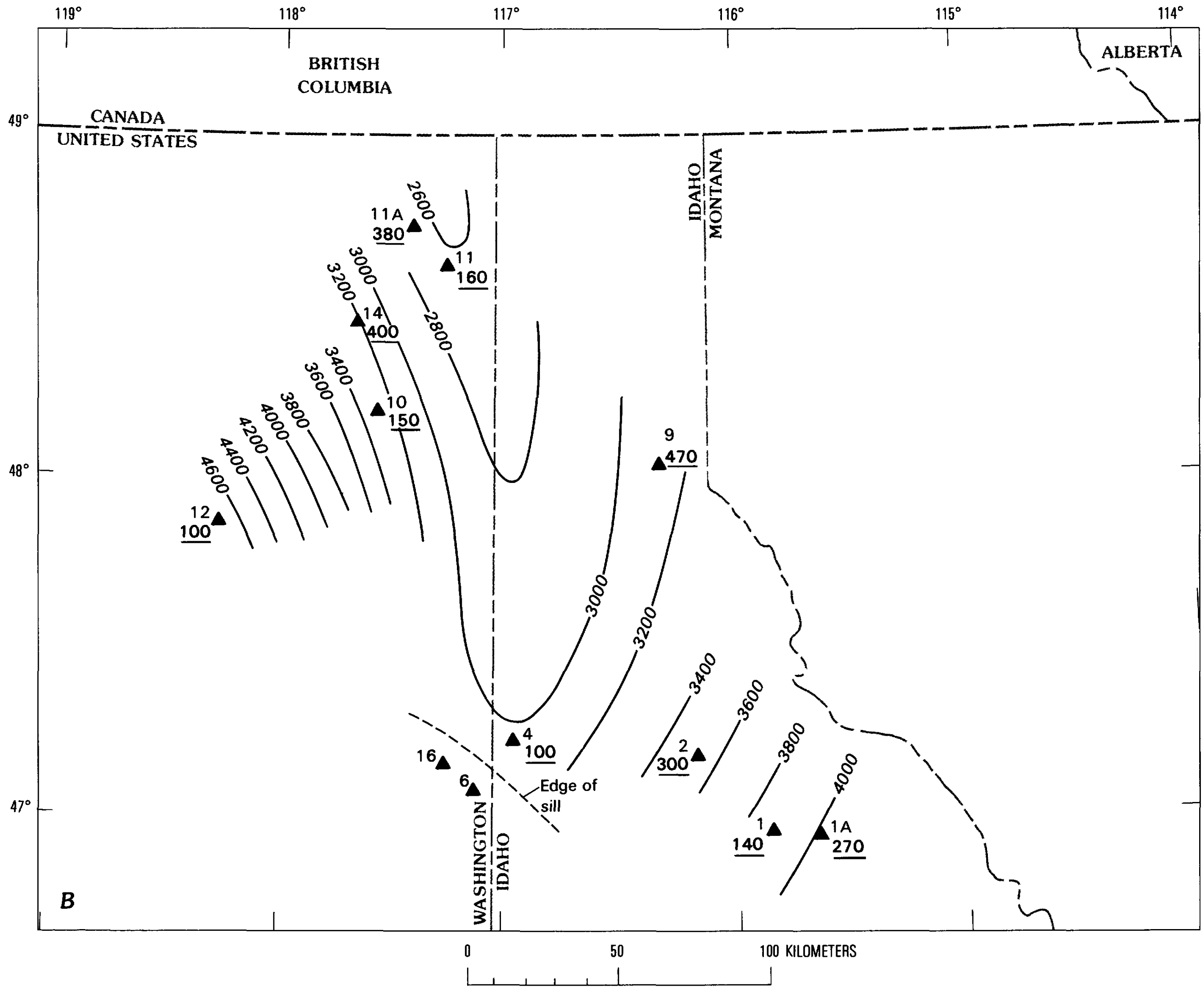


FIGURE 45 (above and facing page).—Contours in meters on top of (*A*) highest sill in section 1 and (*B*) middle sill in section 1. Datum is base of transition member. Triangles are locations of sections used in constructing contours; small numbers identify sections; underlined numbers show thickness of sills in meters at each section. Contour interval, 200 m. Base is palinspastic.

correct, the East Kootenay orogeny probably resulted from convergence of the Siberian and North American cratons. In most of the area of this report the compression produced gentle north- to north-northwest-trending folds (Harrison, 1986). In Canada the compression was more intense and locally produced cleavage (McMechan and Price, 1982, p. 463). In the Coeur d'Alene district northwest-trending folds that formed early in the orogeny were rotated by right-lateral movement along an incipient Lewis-and-Clark line (Hobbs and others, 1965, p. 125, pl. 10). The East Kootenay orogeny was dated by McMechan and Price (1982) at 1,300–1,350 Ma on the bases of the 1,330±45 Ma age of metamorphic biotite in the Prichard Formation (Obradovich and Peterman, 1968), the 1,305±52 Ma rubidium-strontium whole-rock isochron from the Hellroaring Creek stock in British Columbia that intrudes folded rocks of the Aldridge Formation (Leech, 1962; McMechan and Price, 1982, p. 480), and the 1,300–1,350 Ma date for the uppermost Belt inferred by Elston and Bressler (1980) from the polar-wandering curve (Elston, 1984, has more recently inferred an age of 1,250 Ma for the uppermost Belt). Other radiometric ages from the upper part of the Belt Supergroup (Obradovich and others, 1984) would date the orogeny as younger than 1,100 Ma. The final break-up of Siberia and North America was between 555 and 600 Ma (Bond and Kominz, 1984).

The Belt basin may not have been the first such basin



in the northwestern United States. The Yellowjacket Formation (fig. 44) is a basin-fill sequence as much as 6,000 m thick (J.J. Connor, oral commun., 1987) that has been interpreted as a rift deposit that accumulated at about 1.7 Ga (Hughes, 1982; Hahn and Hughes, 1984). The original location of the Yellowjacket basin is not known; the rocks may have been transported many kilometers eastward by thrusting to reach their present position. Additionally, Toth and others (in press) have inferred from lead-isotope ratios in feldspar that the batholithic rocks of the northwestern corner of the Idaho batholith were derived from relatively unevolved material (of Early Proterozoic age) that probably included pelitic sediments in an island-arc setting. This inferred material may have been an extension of the Yellowjacket Formation, or it may have resulted from a separate opening-and-closing event. This is assuming that the evidence for the proximity of the Siberian craton to the west is correct.

If, as suggested above, the Belt basin resulted from an episode of separation and convergence in which little or no oceanic crust was formed, it was similar to many other Proterozoic basins. Dewey and Windley (1981, p. 202–203) wrote that both geologic and paleomagnetic evidence indicate that many Proterozoic orogenic belts were the result of minor separation and resuturing. Commonly in these belts the continental crust was stretched and thinned with but little separation, and the stretching was followed by shortening and stacking of the thinned crust. The Belt basin, then, may have followed a pattern common in the Proterozoic.

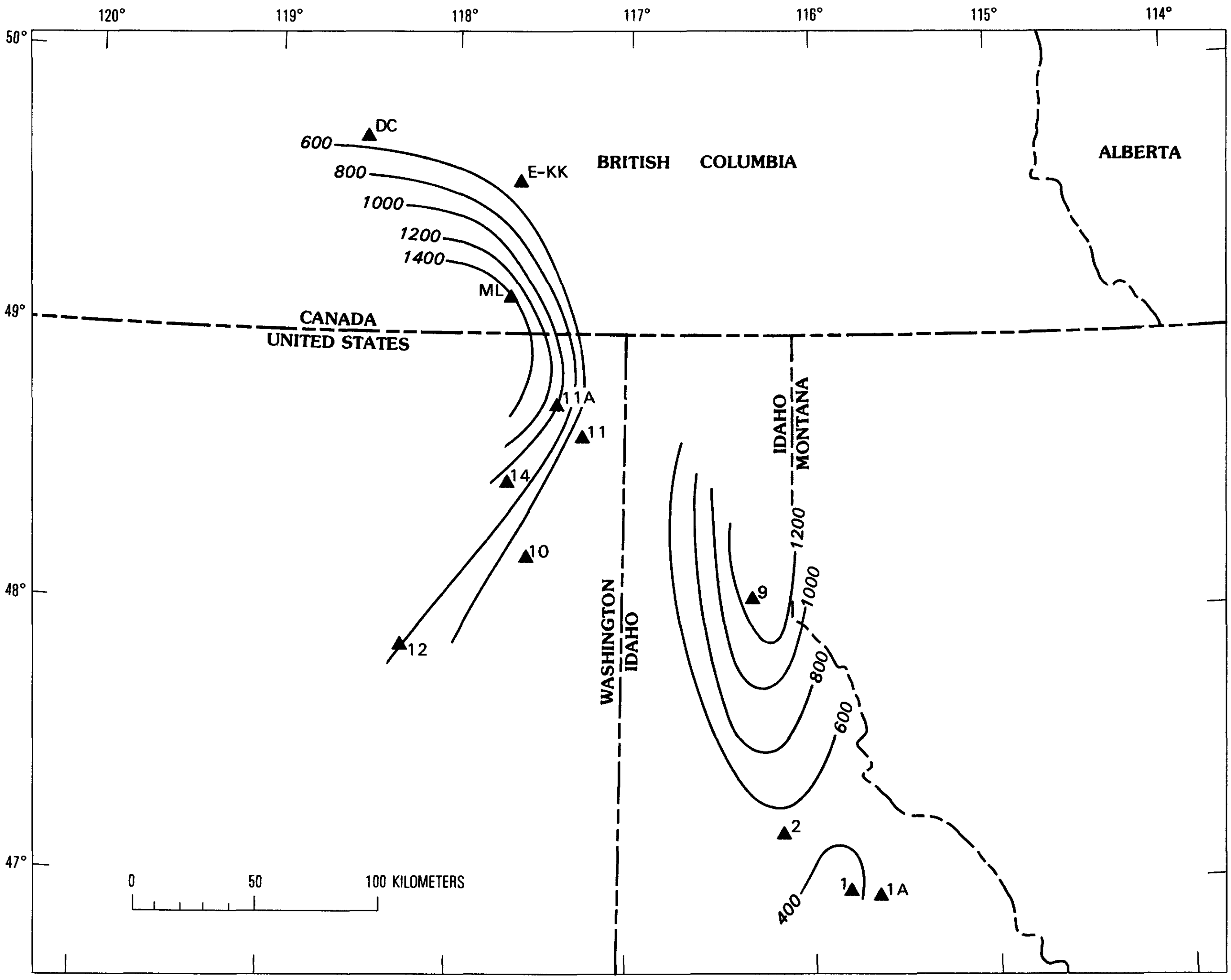


FIGURE 46.—Total thickness of sills in Prichard Formation. Triangles show location of sections used in constructing map. In the United States, small numbers identify sections. In Canada; DC, Dewar Creek (Reesor, 1958); E-KK, Estella-Kootenay King area (Høy, 1979); ML, Moyie Lake area (Høy, 1982). Base is palinspastic.

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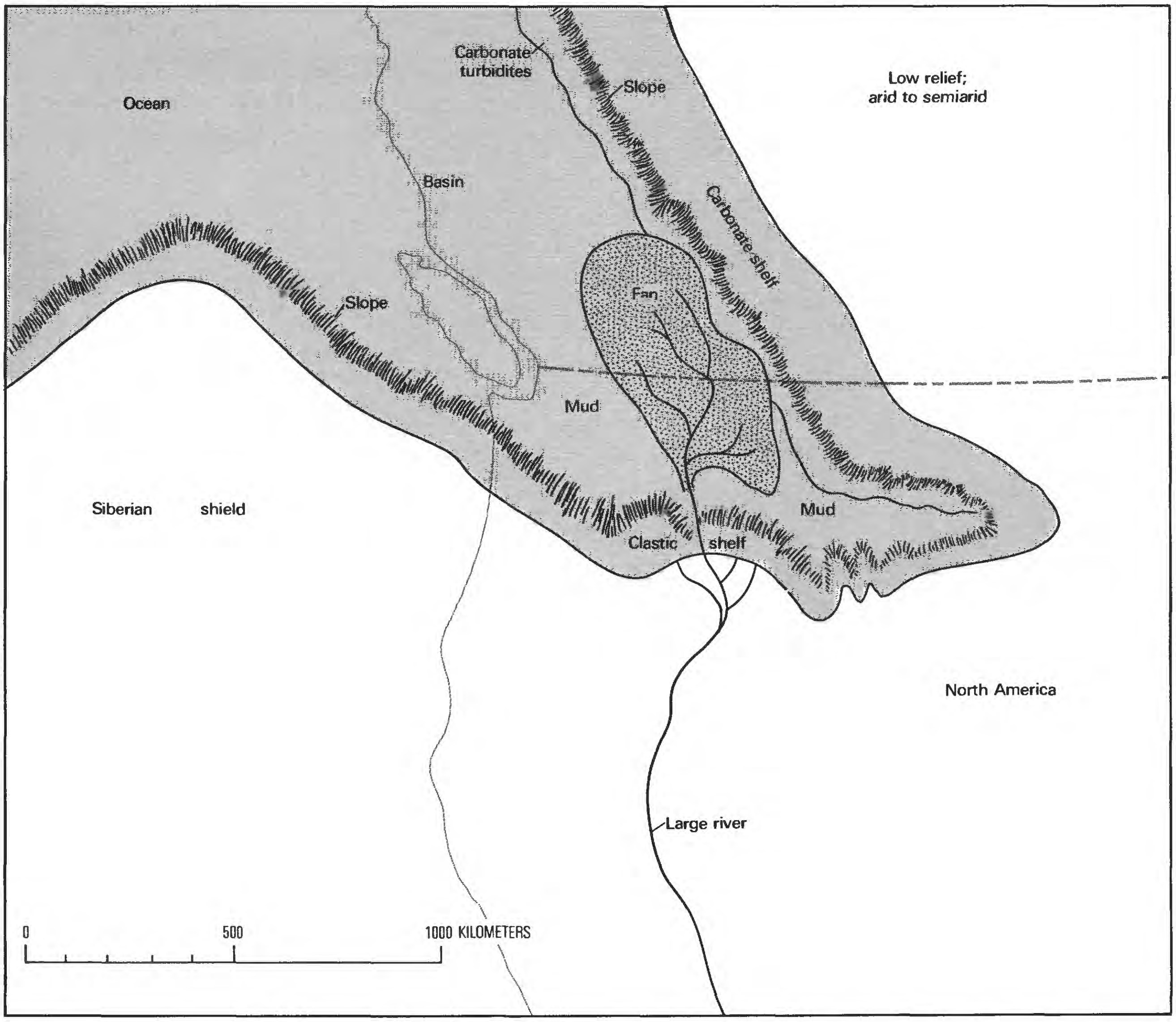

FIGURE 47.—Sketch map showing inferred paleogeography of the Belt basin in Prichard time during second episode of basin filling.

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